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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
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PLAGIOCLASE TWINNING

BY J. D. H. DONNAY*

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ABSTRACT

The French theory of twinning is not based on any genetic consideration; its predictions, notably with respect to the existence and the frequency of twinning, apply to primary and secondary twins alike. In no way can the theory be construed to mean that the age of a given twin is that of the seed crystal. Internal factors control the frequency of the twin if and when the twinning does occur, regardless of its mode of origin.

That external factors, such as thermodynamic conditions, may counteract the effect of the internal factors is recognized. Low index and small obliquity (internal factors) are the prerequisites of twinning; stress and temperature (external factors) only initiate twinning (in secondary twins). If external factors were the ultimate cause of twinning, minerals occurring in the same rock with finely twinned plagioclases would show similar polysynthetic twinning, since they have been subjected to similar conditions.

The width of lamellae in pseudomorphous twins due to replacement or to mutual orientation obviously is not controlled by the composition of the later plagioclase and thus cannot be invoked as an argument against the theoretical predictions as to frequency of twinning.

The undisputed fact that the predicted width of albite-twinning lamellae generally agrees with the observation data is additional proof that external factors do not outweigh internal ones. The effect of external factors, although real, falls short of obliterating the overall statistical picture of the twinning frequency in plagioclases.

INTRODUCTION

In a recent paper Emmons and Gates (1943) have presented evidence to show that polysynthetic twinning of rock-forming plagioclases is mostly of secondary origin, formed late in the growth of the crystals. They ascribe to the French theory of twinning, as stated by Friedel (1904b; 1911; 1926), the idea that "polysynthetic

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twins develop with the crystal's growth" (1943, p. 289) and state that "Donnay (1940) supports this view . . . that the age of a given twin is the age of the seed crystal" (1943, p. 289).

This I believe to be an unfortunate misunderstanding, for which the wording of my paper on the width of albite-twinning lamellae is perhaps partly responsible. The fact is that the French theory of twinning has no bearing on crystallogenesis and offers no solution on the question of age or origin of the twins; it merely recognizes both possibilities, primary and secondary twinnings. Emmons and Gates' conclusions seem to lead to a general indictment of the French theory, inasmuch as these authors are opposed to the primary origin of plagioclase polysynthetic twinning, which they allege is propounded in the theory.

ACKNOWLEDGMENTS

Professor R. C. Emmons kindly consented to read the present paper in manuscript. Although my acknowledgment is not intended to imply that he subscribes to all of its contents, I wish to thank him for his courtesy. It is to be hoped that, when the pressure of war work is relieved, others will join us in the search so that we may understand more about the mechanism and significance of the ubiquitous plagioclase twinning.

FRENCH THEORY INDEPENDENT OF ORIGIN OF TWINNING

The French theory of twinning, which is wholly based on reticular considerations far removed from any crystallogenetic ideas (as to age and origin of twinning), applies equally to either primary or secondary twinning.

Emmons and Gates (1943, p. 289) state that "perhaps, as Donnay suggested, Friedel first placed an age on polysynthetic twins." (I fail to find any such suggestions in my paper.) They quote two passages from Friedel (1904; 1926) to prove his belief in the primary origin of polysynthetic twinning. Friedel, it is true, writes[1] (1926, p. 440):

> "At any time during crystallization, when new particles come to join the crystalline edifice, they can almost indifferently adopt one or the other of the equilibrium positions."

But his very next sentence clearly shows that he also recognizes secondary twinning and that he takes both primary and secondary origins into account:

> "We know, from the existence of mechanical twins[2], that there are even cases where crystalline matter can shift from one of these orientations to another after completion of the crystallization, under mechanical actions."

What Friedel wants to emphasize is not that polysynthetic twinning is primary or secondary, but that the conditions which make twinning possible must prevail continuously during growth, so that twinning may take place at any moment during crystallization (primary twinning) or after crystallization is completed (secondary twinning). Friedel's following sentence brings this out:

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[1] All quotations will be given in translation.

[2] A mechanical twin, in Friedel's terminology, is one that is obtained by gliding, an operation which involves internal reorientations, but which may be considered a "simple shear" as far as the apparent movement is concerned. A celebrated example is that of calcite (Baumhauer's experiment).

"For these polysynthetic twins, and *a fortiori* for mechanical twins, the conditions which make it possible for the twinning position to take the place of the parallel position remain fulfilled during all of the growth."

Friedel, elsewhere, expresses an opinion as to the relative frequencies of mechanical and primary twinnings (1926, p. 425):

"It is far from true that all twins can thus be obtained by a mechanical action exerted on the completely formed (homogeneous) crystal. . . . In the vast majority of cases twinning cannot be obtained by mechanical action, but develops spontaneously during growth."

Now this statement applies to twinning in general and, judging from the scarcity of examples of mechanical twinning that can be obtained in the laboratory, it should not be lightly dismissed. In any case, the foregoing quotations should suffice to prove that Friedel did not exclude secondary twinning as a possible origin of polysynthetic twins.

Emmons and Gates (1943, p. 289) quote a sentence from my paper on the width of albite-twinning lamellae (1940, p. 579), purporting to prove that I support the idea of a primary origin for polysynthetic twinning. In order to explain how a small obliquity would lead to greater ease of twinning (hence, to higher frequency and narrower lamellae), I attempted to show what would happen in the case of primary twinning, thus leading one to infer that I was committing myself to this one genetic hypothesis only. No more than Friedel, however, did I mean to take position in the question of twinning origin, a question which was irrelevant to my problem. Indeed a control of twinning frequency by reticular conditions must hold (by definition, one might say) in secondary as well as in primary twinning, all other things being equal. The reasoning I gave in the case of primary twinning applies almost verbatim to that of secondary twinning. For the sake of completeness I should have added the following (1940, p. 579, after the second paragraph):

"Likewise in the case of secondary twinning (produced after the crystal has completed its homogeneous growth), the width of the lamellae will depend on the ease with which twinning can take place. The frequency of twinning will again be higher, and the lamellae thinner, if the obliquity is small, for slight variations of stress will then suffice to push the particles from one orientation to the other. If the obliquity is large, on the contrary, a change of orientation is not so easy to induce, and the particles will not yield so readily to mechanical twinning, whence fewer and wider lamellae.

"Whatever the factor that initiates twinning formation may be (little understood in primary twins, mechanical in secondary twins), twinning will be more abundant if the crystal is in itself more apt to twin, in other words if the obliquity is small."

RETICULAR CONTROL OF TWINNING NOT INDICATIVE OF PRIMARY ORIGIN

Emmons and Gates write (1943, p. 290):

"If internal factors control the width of albite twin lamellae then the twinning forms with the crystal's growth."

This conclusion does not follow by logical consequence of premise. Internal factors may very well indeed control the width of the lamellae if and when twinning does occur, whether at the beginning, in the course, or at the end of crystallization, or at any time after growth is completed.

Emmons and Gates have offered good evidence that in many cases twinning forms late in the growth of rock-forming plagioclases. They claim that if this evidence is accepted my view (on reticular control) cannot hold. It is clear, from the preceding paragraph, that the acceptance of their evidence (which may well disprove the primary origin of plagioclase twinning in rocks) cannot possibly have any bearing on the validity of the reticular control of twinning.

"CAUSE" OF TWINNING

Emmons and Gates, discussing the cause of twinning, make the following statements (1943, p. 293):

> "We suggest that plagioclase twinning *in general*, in the ordinary rocks referred to here, formed toward the close of the period of crystallization as the result of forces external to the crystals. . . .
> "This viewpoint regards polysynthetic twinning as a gliding effect and denies that it results from the coincidental attachment of molecules to a growing crystal."

Two points must be distinguished: (1) What makes twinning possible at all—the ultimate cause of twinning; and (2) What initiates twinning—its immediate cause. Emmons and Gates do not consider the former. Their remark (1943, p. 289) that

> ". . . twinning in plagioclase is taken quite for granted."

is much to the point in this respect.

The blasting cap that detonates an explosive charge is, in a sense, the "cause" of the explosion, but the immediate cause only. The ultimate cause is the presence of T.N.T. in the charge. No blasting cap of any make will ever cause a sandbag to explode!

An Iceland spar crystal—not twinned—will remain untwinned as long as it is left alone. If I apply a knife to the proper edge, in the proper manner, twinning will result. What is the cause of twinning in calcite? Is it just the mechanical action of the knife blade? Why then do not all species develop twinning under the knife? Because other species may require a more powerful mechanical action to induce twinning? But then, what about the crystalline species that have never been known to develop twinning, under any condition (axinite, chalcanthite)? Could one reasonably maintain that any species is *per se* as apt to twin as any other, and that if some never show twinning it is because the external forces were always either too great or too small? The much abused "explanation" that conditions were not favorable would certainly seem weak in this case, especially if confronted with the facts brought out by Friedel, who observed that in practically every twin law examined[3] some definite geometrical relations are fulfilled (namely, low index and small obliquity).

That forces external to the crystals may "trigger" the twinning is quite evident (since certain species can be made to twin in this fashion in the laboratory); external factors, none the less, are not the ultimate cause of twinning. If they were, then other crystal species in the rock considered would exhibit the same kind of twinning

[3] In his fundamental *Etude sur les groupements cristallins* (1904a), Friedel studied about 150 species or groups of species.

as the plagioclases (like cause, like effect). There must be an "internal" reason why various minerals react differently to similar environmental conditions; why certain species are apt to show twinning and others not; why some, like oligoclase, usually twin more readily than others, like anorthite. This reason has been found in the reticular geometry of the twinned crystals.

PSEUDOMORPHOUS TWINS NOT DUE TO "TWINNING"

MUTUAL ORIENTATION

Emmons and Gates (1943, Pl. 3, fig. 3, and p. 294) describe, occurring in a gabbro, a plagioclase crystal An_{55} with relatively few lamellae, which extends into a granite dike as An_{18}, without any modification in the width of the twinning lamellae. This they consider (1943, p. 295)

> "a striking refutation of the claim that composition controls the width of twin lamellae."

Is it necessary to point out that the secondary growth of the plagioclase An_{55}, by addition of material from another solution (An_{18}) after twinning had formed (in the gabbro), could not be expected to behave otherwise? The orientation of incoming particles is controlled by that of the lamella on which they attach themselves. This is an example of the well-known phenomenon of mutual orientation[4], which has nothing to do with twinning. Indeed, although the An_{18} extension of the An_{55} plagioclase does show "twinning lamellae," these lamellae of An_{18} have not in any sense been produced by "twinning." This, incidentally, is exactly the view Emmons and Gates take of the phenomenon (1943, p. 294):

> "When a feldspar is once twinned, later growths of feldspar on the same crystal usually continue the lamellation."

The same idea is also expressed elsewhere (1943, p. 292):

> "Crystal enlargements normally continue the twinning already formed."

The case is clear. According to the French theory of twinning, the width of the lamellae in the An_{18} increment should be controlled by the twin obliquity in An_{55} and should be moderately large. Judging from Emmons and Gates' excellent photomicrograph (1943, Pl. 3, fig. 3), it appears to be so.[5]

REPLACEMENT

Pseudomorphous twinning lamellae may also arise from replacement. The width of the lamellae, in such cases, cannot be controlled by the anorthite content of the

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[4] Two (or more) different species can concur to the building of one and the same crystalline edifice provided (among other conditions) that some cell (not necessarily the unit cell) of the one species be approximately of the same shape and size as a cell of the other species. The most famous example is perhaps that of the various alums, which can be made to grow (from successive solutions of different compositions) into what appears to be a single crystal, but actually is a zoned edifice constituted by successive crystals encasing one another. The phenomenon of mutual orientation is closely akin to that of isomorphous syncrystallization, which, however, gives rise to "mixed crystals."

[5] The fact that the "ocular times objective" magnifications given by Emmons and Gates (1943, p. 290, footnote) are not the actual magnifications of the illustrations makes it difficult to compare the frequency of the twinning from figure to figure. The width of lamellae is susceptible of a rigorous, quantitative definition—it is inversely proportional to the frequency of the polysynthetic twinning, that is to say to the number of lamellae per unit length (provided the plane of the thin section is perpendicular to the twin plane and the unit length considered is taken perpendicular to the trace of the twin plane).

replacing plagioclase, any more than the "crystal form" of martite octahedra could be correlated with their hematite composition. The width of the lamellae should be governed by the twin obliquity in the original plagioclase.

CORRELATION BETWEEN WIDTH OF LAMELLAE AND ANORTHITE CONTENT

To the observations of Dufresnoy, Delafosse, Des Cloizeaux, Fouqué and Michel-Lévy, Tschermak, and A. de Lapparent, on which I drew to substantiate the predictions of the theory (1940, p. 584–585), it is a pleasure to add the considered opinion of Emmons and Gates (1943, p. 295):

"It is well recognized that *in general* calcic plagioclase and very sodic plagioclase typically have broad lamellae and that plagioclase of composition about oligoclase has relatively narrow lamellae."

Their statement acquires even more weight in view of the long experience they have had in hunting wide lamellae in plagioclases (1943, p. 296):

"A careful search over a period of years for feldspar material coarsely enough twinned to make an overall separation of adjacent lamellae possible has been most discouraging. Two such types of plagioclase have been found. . . . They are the very calcic plagioclase . . . , and the pegmatite specimens such as the coarsely twinned albites of Amelia Court House, Virginia . . . in oligoclase and andesine, . . . the lamellae are hopelessly narrow."

The facts thus appear undisputably established. The overall statistical picture is that the frequency of polysynthetic twinning is highest in the oligoclase region (narrow lamellae) and lowest toward both end members of the series (wide lamellae, dual twinning, to single crystals). These facts are predicted, along with many others, by the French theory of twinning. According to Emmons and Gates' argumentation, this remarkable agreement can only be explained by a whim of chance, a coincidence.

Now it also happens, and these are also facts: (1) that the basicity of the host rock controls (to a large extent) the anorthite content of its plagioclase; (2) that viscosity is low in basic and very acid rocks. Hence the explanation that the broad lamellae (or single crystals) found in anorthites and very sodic albites are due to the low viscosity of the host rock. No flaw can be found in this reasoning insofar as it applies to rocks, but any generalization of the explanation would be hazardous. Will it account for the results obtained by Fouqué and Michel-Lévy in their plagioclase syntheses?

CONCLUSIONS

The reticular conditions (low index, small obliquity) which make it possible for twinning to occur at all, also account, all other things being equal, for the relative frequencies (or width of lamellae) of polysynthetic albite twinning. These "internal factors" exert their influence whether the twinning is of primary (spontaneous) or secondary (mechanical) origin.

The effect of "external factors" (thermodynamic conditions, in general) may counteract the reticular influence. It bids fair to be of sufficient magnitude to yield valuable information, in particular cases, on the conditions to which the rock must have been subjected. Inferences, however, must be drawn with caution. In order

properly to appraise the action of the (varying) external factors on a given polysynthetic twin, one should first estimate the effect to be expected from the (constant) internal factors. If an oligoclase, for instance, displays wide lamellae, contrary to what should be expected from its small obliquity, then it is quite reasonable to look for an explanation in the environmental conditions. If, on the other hand, the oligoclase shows its usual "hopelessly narrow" lamellae, what justification is there for ascribing the fine lamellae to the high viscosity of the rock, rather than to the small obliquity of the twin which suffices to account for them in synthetic oligoclase?

By and large, the effect of the external factors, real though it may be, hardly affects the general statistical picture of agreement between facts and theory insofar as the width of albite-twinning lamellae is concerned. In this sense, internal factors of twin control may be said to outweigh external ones.

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PERMIAN TYPE SECTIONS IN CENTRAL NEW MEXICO

BY C. E. NEEDHAM AND ROBERT L. BATES

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ABSTRACT

Standard usage has established four Permian formations in central New Mexico above a thin unnamed basal Permian limestone. In ascending order these are: the Abo formation — type locality in Abo Canyon, Valencia and Torrance counties, New Mexico; Yeso formation — type locality near Mesa del Yeso, Socorro County; Glorieta sandstone — type locality in Glorieta Mesa, San Miguel County; and San Andres limestone — type locality in Rhodes Canyon, San Andres Mountains, Socorro County. The type sections for these formations have never been adequately designated, and the formations at the type localities have not been sufficiently delimited and described. This paper attempts to designate the type section, define the limits, and describe the characteristic lithology of each formation at its type locality.

GENERAL STATEMENT

The basal Permian formation in central New Mexico is a thin marine limestone of Wolfcamp age, which will be described and named by M. L. Thompson. Above this formation are four others—in ascending order: the Abo, Yeso, Glorieta, and San Andres formations. More precise knowledge concerning these four units is now needed, not only by surface stratigraphers, but perhaps more urgently by subsurface

geologists who first encounter the formations in the depths of the great West Texas-New Mexico Permian Basin and wish to add surface information to that secured from cuttings and cores. Some of the most baffling problems of Permian Basin geology remain unsolved because information as to outcrop, character, and extent of the Permian formations of central New Mexico has not been available. It is the purpose of this report to supply some of this information.

The Abo, Yeso, Glorieta, and San Andres formations all have their type sections in central New Mexico. As originally described, these type sections were inadequately designated as to location and only very loosely defined as to lithology. The writers have therefore revisited and redescribed the type section of each formation. No attempt has been made to rename a formation or to change the type locality; on the contrary, the aim was always to stay as close as possible to the originally intended type locality and to describe as nearly as possible the originally indicated rock unit. Brief remarks on such matters as thickness, distribution, and correlation of each formation have been added to its redescription. Although the report has been written by the junior author, the conclusions are those of both authors and are based on joint field work.

ABO FORMATION

ORIGINAL DESCRIPTION

Lee (1909, p. 12) gave the name Abo sandstone to a succession of dark-red coarse sandstones, in part conglomeratic, exposed in Abo Canyon at the south end of the Manzano Mountains. Maximum thickness was given as 800 feet. Lee stated that the Abo was underlain unconformably by the Magdalena limestone (Pennsylvanian). His definition of the upper boundary is not clear. He termed the Abo the basal member of the Manzano group, which contained, in ascending order, the Abo sandstone, the Yeso formation, and the San Andres limestone.

Reference to Lee's discussion of the exposures in Abo Canyon shows that no specific type locality was set up and that no type section was described. Furthermore, several thin fossiliferous limestones that Lee described as lying near the base of the Abo are now known to belong in the underlying Pennsylvanian. Lee's conclusion that shale in the Abo is subordinate in amount to sandstone and conglomerate is also in error, as reference to the following measured section will show.

REDESCRIPTION

Exposures of the Abo formation in Abo Canyon are relatively poor. Sandstone and arkose beds are prominent, but many of the intervening shales are covered. The gentle eastward dip makes it necessary to traverse a considerable horizontal distance in order to cover the entire sequence of beds. For these reasons, the section measured by the writers contains a number of covered intervals and is some 4 miles long. It is described as follows: From the base of the formation in sec. 32, T. 3 N., R. 5 E., Valencia County, just north of U. S. Highway 60, through sec. 33, into sec. 2, T. 2 N., R. 5 E., Torrance County; thence to secs. 35 and 36, T. 3 N., R. 5 E., to the upper limit of the formation at the top of the hill just east of the road to Abo Ruins in sec. 25, T. 3 N., R. 5 E., Torrance County. The base of the formation lies about 1 mile

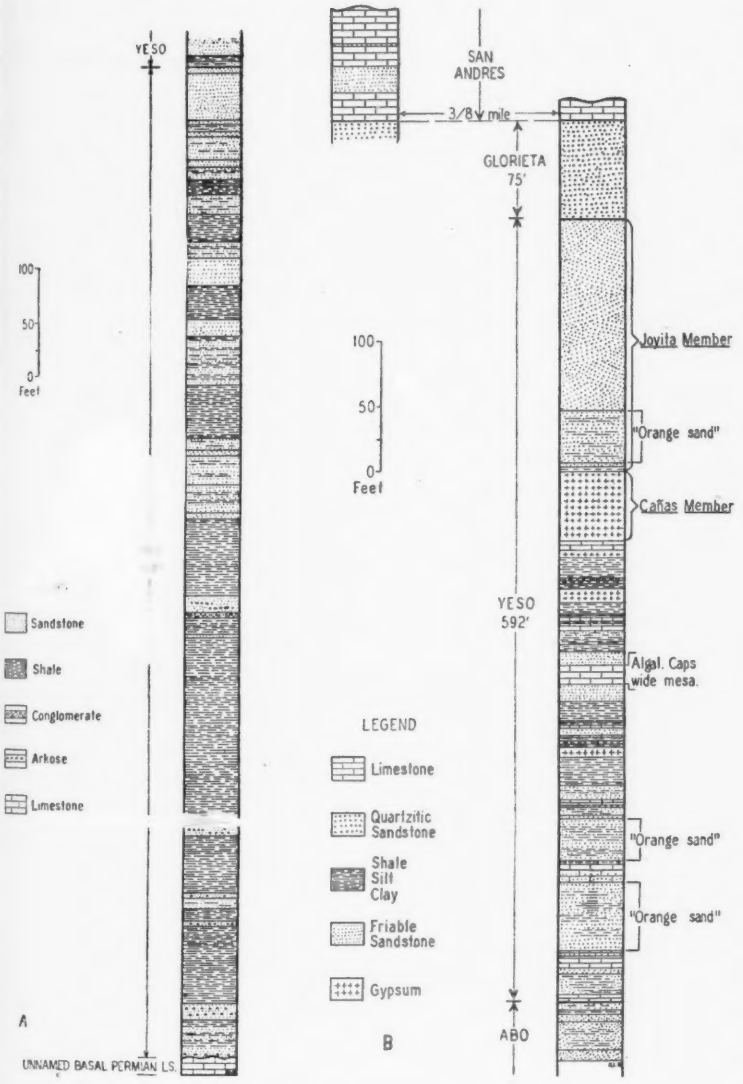

FIGURE 1.—*Columnar sections*

(A) Columnar section of the Abo formation at the type locality.
(B) Columnar section of the Yeso and Glorieta formations at the type locality of the Yeso.

northwest of the village of Scholle, and its top about 2 miles west-northwest of the village of Abo.

Following is the section measured by the writers at the above locality. (*See also* Figure 1A.)

Interval	Description	Thickness (in feet)
Abo formation		
40.	Sandstone, white, one massive bed; upper part weathers limy. The top of this sandstone is an erosion surface. It is considered the top unit of the Abo on the basis of a section measured on a country road 7 miles west of the town of Mountainair, in which section the same sandstone is found overlain by typical Yeso beds	6.0
39.	Sandstone, pink, medium-grained, in beds of medium thickness	43.0
38.	Sandstone, light-colored, coarse	3.0
37.	Shale, pink, sandy, with sandstone parting	7.0
36.	Sandstone, light-colored, coarse	5.5
35.	Sandstone, pink and buff, with shale partings	21.0
34.	Sandstone, red	7.0
33.	Sandstone, pink, shaly; and sandy shale	12.0
32.	Shale, red	3.0
31.	Sandstone, light pink, medium-grained	2.0
30.	Shale, pink	8.0
29.	Sandstone, buff, thin-bedded and shaly, fine-grained	18.0
28.	Shale, red	25.0
27.	Sandstone and shale, buff	15.0
26.	Sandstone, pink, medium-grained, thin-bedded in lower part, massive in upper	25.0
25.	Shale, buff, partly covered	32.0
24.	Sandstone, pink, medium- to heavy-bedded; caps mesa	16.0
23.	Sandstone and shale, pink and red, interbedded; partly covered	45.0
22.	Shale, red, partly covered	48.0
21.	Sandstone, red, shaly	12.0
20.	Sandstone, red, medium-bedded, medium-grained, cross-bedded; weathers blocky. This bed and No. 21 cap a wide prominent cuesta	4.5
19.	Shale, yellow-gray	0.5
18.	Sandstone, red, fine, shaly, extremely cross-bedded	27.0
17.	Shale and sandstone, red, partly covered	26.0
16.	Sandstone, red, medium-bedded, medium-grained, cross-bedded	6.0
15.	Shale, red, partly covered	70.0
14.	Sandstone, red, coarse, cross-bedded; carries conglomerate lenses	16.0
13.	Conglomerate; limestone pebbles in red sandstone	5.0
12.	Shale, red, partly covered	16.0
11.	Sandstone, red, slabby	3.0
10.	Mostly covered; some red sandy shale	172.0
9.	Sandstone and arkose, dark-red, coarse	10.0
8.	Covered; probably red shale	52.0
7.	Sandstone, red, arkosic, coarse; forms ledge	6.0
6.	Sandstone, red, shaly, arkosic	10.0
5.	Covered; probably red shale	11.0
4.	Sandstone, red-brown, coarse, arkosic, conglomeratic	6.0
3.	Covered; probably red shale	70.0
2.	Arkose and sandstone, red-brown, thick-bedded, with fragments of limestone	15.0
1.	Sandstone and shale, interbedded, partly covered	35.0
	Total thickness of Abo formation	914.5

Unnamed basal Permian limestone.

ADDITIONAL NOTES

Lithology: In its type section the Abo consists of about 60 per cent red shale and about 40 per cent sandstone, arkose, and conglomerate. This ratio holds approximately true for most exposures in central New Mexico. It is therefore incorrect to refer to the formation as the Abo sandstone.

The Abo contains no limestones or marine fossils. Its continental origin is indicated by red color and clastic character; casts of halite crystals; mud cracks; ripple mark, both wave and current; very extensive cross-bedding; bones and tracks of land vertebrates; and plant remains.

THICKNESS: Few figures are available. The thickness of 915 feet at the type section is the maximum known to the writers, but the formation may well be thicker at other places. Darton (1922, p. 254) states that the Abo is 600 to 700 feet thick at the head of Zuni Canyon in the Zuni Mountains. About 630 feet has been measured in the La Joyita Hills 18 miles north of Socorro, where a few feet may be missing against a fault.

DISTRIBUTION: The Abo formation is known from the Zuni Mountains in Valencia and McKinley counties eastward to the region around Las Vegas, and from the Chama Basin in southern Rio Arriba County southward in disconnected outcrops to the San Andres and Sacramento mountains. Farther south, it changes in facies into the Hueco limestone in the Hueco Mountains of Texas.

CONTACTS: The lower contact of the Abo is an unconformity, as the formation rests on beds ranging in age from earliest Permian in the region of the type locality down to the pre-Cambrian in the Zuni Mountains. The upper contact with the Yeso appears in many places to be gradational. Especially in the northern part of the State, the change in color and lithology from typical Abo sandstones and shales upward into characteristic Yeso pink sands and silts is gradual, and the contact is difficult to place. Farther south, for example at the Yeso type locality, a thin basal limestone of the Yeso separates that formation from the Abo redbeds.

PHYSIOGRAPHIC EXPRESSION: In most places the alternating nonresistant shales and resistant sandstones of the Abo give rise to cuestas, forming a valley-and-ridge topography. The formation produces little soil and supports only a thin growth of vegetation, so that its characteristic bright-red color can be distinguished for long distances.

CORRELATION: To the south the Abo grades laterally into the Hueco limestone, and to the north it is believed to be equivalent to the Rico limestone of Colorado. In the Grand Canyon region of Arizona it is probably represented by the Supai formation. The Abo is a part of the Wolfcamp series of early Permian age.

YESO FORMATION

ORIGINAL DESCRIPTION

Lee (1909, p. 12) named the Yeso and designated it the middle formation of the Manzano group. The name is derived from the Mesa del Yeso, a small tableland 12 miles northeast of Socorro. Lee recognized Yeso beds in the slopes of this mesa and stated that the formation is "typically exposed" there. However, he made no further mention of this exposure. The section he described in text, columnar section, and faunal list lies some 2 miles southeast of the Mesa del Yeso. Considering that the exposures at the mesa are poor and incomplete, the writers feel that Lee was wise to choose the section 2 miles away. The section given below was measured in the immediate vicinity of Lee's section, and some parts are doubtless identical with his.

Lee's description of the Yeso is unsatisfactory because:

(1) The type locality is not completely designated.

(2) Lee's section is measured in such general terms as to be unsuitable for present stratigraphic needs.

(3) At the top of Lee's columnar section, and included by him in the Yeso formation, is a 75-foot limestone the top of which is an exposed erosion surface. This limestone is now known to be basal San Andres.

(4) Below this bed, Lee also included in the Yeso 200 feet of sandstone, part of which is now separated as the Glorieta sandstone.

(5) A limestone layer at the base of the Yeso is described as highly fossiliferous (32 species were collected from it) and as 50 feet thick. Such a layer is not believed to occur at or near the base of the formation. Although the writers place the base of the Yeso at the base of a limestone bed, this bed is only 7 to 10 feet thick and is barren of fossils. A prominent exposure of San Andres limestone is found three-eighths of a mile west of the base of the Yeso type section as herein redescribed, the position of which is the result of faulting and the relations of which are obscure without detailed investigation. As this San Andres stratum is about 50 feet thick and contains many fossils, it is thought that Lee may have mistakenly included it in the basal Yeso.

REDESCRIPTION

The base of the section measured by the writers is 11.2 miles N. 46° E. of Socorro at the point on the Socorro quadrangle where its eastern edge is intersected by the 34°10′ parallel. The section was measured northeast from this point, in canyons and cuesta escarpments in secs. 4 and 5, T. 2 S., R. 2 E., and in sec. 33, T. 1 S., R. 2 E., Socorro County. The top of the section is $2\frac{1}{4}$ miles southeast of the Mesa del Yeso, on a high isolated butte that can readily be recognized because of a prominent double erosion pillar on its west slope.

The following section was measured. As the Glorieta sandstone is well exposed, it is included. (*See also* Figure 1B.)

Interval	Description	Thickness (in feet)
	San Andres formation (Upper part eroded)	
59.	Limestone, gray, nonfossiliferous	15.0
	Glorieta sandstone	
58.	Sandstone, white to light-gray, quartzitic; weathers angular and blocky	75.0
	Yeso formation	
57.	Sandstone, pink in lower part, white to light-gray in upper part; cross-bedded; weathers to rounded forms	145.0
56.	Sandstone, red and pink, thin-bedded and shaly; "orange sand"	40.0
55.	Sandstone, white to light gray	3.0
54.	Silt and clay, pink and gray	4.5
53.	Gypsum, with pink silt partings	52.0
52.	Limestone, gray	7.0
51.	Gypsum	5.5
50.	Silt and clay, varicolored	15.0
49.	Gypsum	1.0
48.	Gypsum and pink silt	2.5
47.	Silt, varicolored	3.0

Interval	Description	Thickness (in feet)
46.	Gypsum	10.7
45.	Silt and clay, pink and gray	8.0
44.	Limestone, gray	1.0
43.	Silt, gray and pink	1.5
42.	Limestone, gray, with silt parting	3.0
41.	Silt and shale, calcareous and gypsiferous	1.5
40.	Limestone, gray, dense	2.0
39.	Silt, pink	0.5
38.	Limestone, gray, dense, thin-bedded; caps low mesa	3.0
37.	Silt, yellowish; sandy at top	11.0
36.	Silt, pink, gypsiferous	6.0
35.	Sandstone, red to pink, thin-bedded, silty	10.5
34.	Limestone, dolomitic, gray, dense, massive; weathers rough; makes capping of broad mesa; top bed carries algal remains	14.5
33.	Sandstone, yellow at top and base, red in middle part; weathers to rounded forms	13.0
32.	Silt and shale, pink; some gypsum	15.0
31.	Limestone, dark-gray	0.8
30.	Silt, pink to white	2.0
29.	Limestone, dark, massive	2.3
28.	Sandstone, gray, calcareous; weathers rounded	5.0
27.	Limestone, gray	1.5
26.	Silt, orange	3.0
25.	Gypsum	2.5
24.	Silt, orange	2.0
23.	Gypsum	7.0
22.	Shale and silt, red to orange	22.0
21.	Sandstone and shale, light-colored; interbedded	11.0
20.	Limestone, light gray at base, black at top; thin- to medium-bedded	4.0
19.	Sandstone, buff, shaly	1.5
18.	Sandstone and shale, light-colored, thin-bedded	6.0
17.	Sandstone, pink and white, friable	3.0
16.	Silt and sandstone, orange, thin-bedded to massive	32.0
15.	Silt and gypsum, white	3.0
14.	Limestone, gray, medium-bedded; very sandy in lower 8 feet, silty above	13.0
13.	Sandstone and silt, orange, massive, fine; gray at top	52.0
12.	Limestone, gray, dense, medium-bedded; weathers rough and pitted	1.5
11.	Sandstone, yellow, silty, thin-bedded	2.5
10.	Covered	1.0
9.	Limestone, gray, dense, medium-bedded, crystalline	7.3
8.	Sandstone, yellow, soft, silty	5.4
7.	Silt and sandstone, pink to buff, fine; poorly exposed	15.0
6.	Limestone, gray, dense, thin-bedded, nonfossiliferous, cherty	6.7
	Total thickness of Yeso formation	592.7
	Abo formation (Only upper beds measured)	
5.	Sandstone, gray, shaly	5.5
4.	Shale, tan to white, sandy, calcareous	4.0
3.	Sandstone, red to buff, cross-bedded, massive	5.0
2.	Shale, red, sandy, cross-bedded	22.0
1.	Sandstone, dark-red and buff, cross-bedded; changes color laterally; weathers to rounded ledges	8.0

MEMBERS

The Yeso formation may be subdivided into at least four units. The lowest is a general zone of clastic material, characterized by abundant pink or orange sandstone. Next above is a thick succession of interbedded thin limestones, gypsums, silts, and sandy shales, with progressively more limestone and gypsum toward the southeast. This unit is informally referred to as the middle evaporites. In Otero County a prominent bed of limestone is present in the lower part of this unit, and in the vicinity

of Bent, northeast of Tularosa, and in Nogal Canyon to the south it is 85 to 100 feet thick. Neither the lower clastic zone nor the middle evaporites are thought distinct enough as to upper and lower boundaries or consistent enough in lithology to merit a name.

Above the middle evaporite unit is a thick bed of gypsum which is here named the Cañas member of the Yeso formation. The name is taken from the Lomo de las Cañas, a range of hills 9 miles due east of Socorro, at the north end of which the gypsum member is well developed and exposed. The Cañas member is 96 feet thick at the Lomo de las Cañas, 52 feet in the Yeso type section, 80 feet in southern Chupadera Mesa, and 115 feet at Bent, Otero County. It disappears a short distance north of the Yeso type section, as it is not found in a completely exposed Yeso section in the La Joyita Hills 18 miles north of Socorro. In most places the Cañas gypsum is white and pure, although thin silt partings are not unknown. V. C. Kelley (personal communication, 1942) states that at Jones Camp, in the south part of Chupadera Mesa, the Cañas member contains a number of thin limestone beds. At the Lomo de las Cañas, Chupadera Mesa, and Bent, the Cañas gypsum is underlain by a limestone 10 to 20 feet thick which carries distinctive almond-shaped organic remains of unknown affinities. The Cañas member is overlain by the Joyita sandstone member.

Wherever the Yeso has been studied by the writers, its uppermost unit consists of a soft cross-bedded sandstone to which the name Joyita member of the Yeso formation is here given. The name is taken from a well-exposed section in the canyon of a west-flowing tributary to the Rio Grande in the La Joyita Hills 18 miles north of Socorro and about 4½ miles south of the village of La Joya. At this place the Joyita member is 160 feet thick and consists of pink, orange, and yellow thin-bedded sandstone that weathers to rounded ledges. The orange color and the manner of weathering distinguish the Joyita member. It is 185 feet thick in the Yeso type section, 120 feet at Chupadera Mesa, 105 feet at Bent, and is believed to be 105 feet thick in the Texas Production Company No. 1 State-Wilson, a dry hole on the Dunken Dome in southwestern Chaves County. It is underlain by limestones of the middle evaporites or by the Cañas gypsum, and overlain by the quartzitic, resistant, blocky Glorieta sandstone. In a few places, as at the Lomo de las Cañas, it is difficult to place the boundary between the Joyita member of the Yeso and the Glorieta sandstone, as the lithology of the two units is very similar.

ADDITIONAL NOTES

THICKNESS: The Yeso formation thickens toward both the northwest and the southeast from the type section. According to Parry Reiche (personal communication, 1942) some 900 feet is exposed in the Zuni Mountains in Valencia and McKinley counties. A well-exposed section at Rhodes Pass, in the San Andres Mountains 75 miles south of the type section, also shows 900 feet. At Jones Camp in Chupadera Mesa 750 feet is present. Farther east, in the northern Sacramentos, the writers have measured 1050 feet, with an estimated 100 feet concealed between the base of the exposed section and the top of the Abo. In the deep test well on Dunken Dome, on the east slope of the Sacramento Mountains, the Yeso is nearly 2000 feet thick.

DISTRIBUTION: The Yeso crops out in the Zuni Mountains, in Glorieta Mesa, and

in most of the mountain ranges of central New Mexico. In addition, it is known to extend far to the south and east in the subsurface of the Permian Basin.

FOSSILS: In its typical development in central New Mexico the Yeso contains no distinct fossils other than definite algae which are found in a limestone near the middle of the formation. The peculiar almond-shaped organisms found in the limestone beneath the Cañas gypsum member are thought to be algae, although according to J. Harlan Johnson (personal communication, 1942) they are unlike any that have so far been described.

In the Yeso limestones of the southeastern part of the State, brachiopods and gastropods are found in some abundance. The Yeso fauna has not been described in detail.

CONTACTS: In the type section the Yeso is sharply set off from the underlying Abo redbeds by a thin prominent limestone. This is also true in the La Joyita Hills and seems to be representative of the condition in central New Mexico. To the north and northwest, however, the limestone disappears, and the lower clastic phase of the Yeso is in contact with the dark red shaly sandstone of the Abo. A disconformity is thus postulated at the base of the Yeso formation. The Glorieta sandstone appears to overlie the Yeso conformably, with a plain lithologic break in most places and a gradual transition in others.

PHYSIOGRAPHIC EXPRESSION: Because it is overlain by the resistant Glorieta and San Andres formations, the Yeso as a rule does not crop out in wide areas but rather in narrow belts along mountain escarpments. The thin limestones are resistant, forming ledges and capping small cuestas. The sandstones weather to bare rounded ledges with a tendency toward exfoliation. The other lithologic types make steep slopes on canyon walls and escarpments.

CYCLICAL DEPOSITION: In parts of the Yeso formation deposition appears to have been cyclical or rhythmic. The presence of eight thin limestone beds in the lower two thirds of the formation, separated by sand and silt beds and in places by gypsum, shows that conditions of sedimentation returned repeatedly to the production of carbonate material. Cyclical deposition in the Cañas gypsum member is suggested by the presence at some places of at least four thin and apparently persistent beds of silt, and at other places of thin limestone layers. Further work is necessary to demonstrate the significance of these evidences of rhythmic deposition.

STRATIGRAPHIC RANK: In his original description Lee made the Yeso a formation. Darton (1922, p. 181) reduced it to the position of a member of the Chupadera formation, a term he introduced to include beds of Yeso and San Andres age. The term Chupadera, however, has been abandoned by the U. S. Geological Survey (King, 1942, p. 687). The Yeso is now considered a formation of the Manzano group.

CORRELATION: The writers consider that the Yeso belongs in the lower part of the Leonard series. It is thought to be equivalent to the lower part of the Bone Spring limestone of the Delaware basin and probably to the lower DeChelly sandstone of northeastern Arizona. It was termed Nogal formation by Nye (1933, p. 70–76), but this name is now abandoned.

GLORIETA SANDSTONE

GENERAL STATEMENT

The name Glorieta was first used by Keyes (1915, p. 2, 7), who applied it to the main body of the Dakota sandstone (Cretaceous) around the south end of the Rocky Mountains. Although Keyes gave no type locality, presumably the sandstone was named from Glorieta Mesa in Santa Fe and San Miguel counties, New Mexico, or for the town of Glorieta at the north end of the mesa. Cretaceous formations do not crop out at either of these places. Common usage has determined the Glorieta to be the prominent sandstone, well developed and exposed on Glorieta Mesa, that separates the Yeso and the San Andres formations.

The Glorieta was first designated Permian by Hager and Robitaille (1919). Baker (1920, p. 111, 118–119, 126) stated that it was the basal member of the Upper Triassic, but, as pointed out by Rich (1921, p. 295–296), it is apparent that Baker confused the Glorieta with the Santa Rosa sandstone of Triassic age, of which erosional remnants lie on Glorieta Mesa.

DESCRIPTION

The writers designate as the type section of the Glorieta sandstone a good exposure in Glorieta Mesa 1 mile west of the village of Rowe, San Miguel County, New Mexico. The section lies in the south-central part of T. 15 N., R. 12 E. It is plainly visible from U. S. Highway 85 and from the Santa Fe Railway.

As other beds besides the Glorieta are well exposed, they are included in the following section. (*See also* Figure 2A.)

Interval	Description	Thickness (in feet)
	Santa Rosa sandstone (Upper part eroded)	
9.	Sandstone, light colored, massive, coarse; caps mesa	14.0
8.	Sandstone, pink, shaly, very thin-bedded	64.0
7.	Sandstone, light to pale pink, with some conglomerate; cross-bedded, massive; weathers rounded; makes cliff	40.0
	Unconformity?	
	Triassic or Permian	
6.	Sandstone and sandy shale, orange and pink, thin-bedded and slabby	64.0
	Unconformity?	
	San Andres formation	
5.	Limestone, gray to buff, thin-bedded, partly sandy	20.0
	Glorieta sandstone	
4.	Sandstone, white to gray, medium coarse, quartzitic, in beds 2 to 6 feet thick; weathers pale yellow; forms cliff	116.0
3.	Sandstone, buff to white, thin-bedded, soft; poorly exposed	20.0
	Total thickness of Glorieta sandstone	136.0
	Yeso formation	
2.	Sandstone and silt, orange to pink, uniform; in massive beds that weather rounded and make steep slopes	301.0
	Abo formation (Only upper beds measured)	
1.	Sandstone, white, massive, clean, medium to coarse	5.0

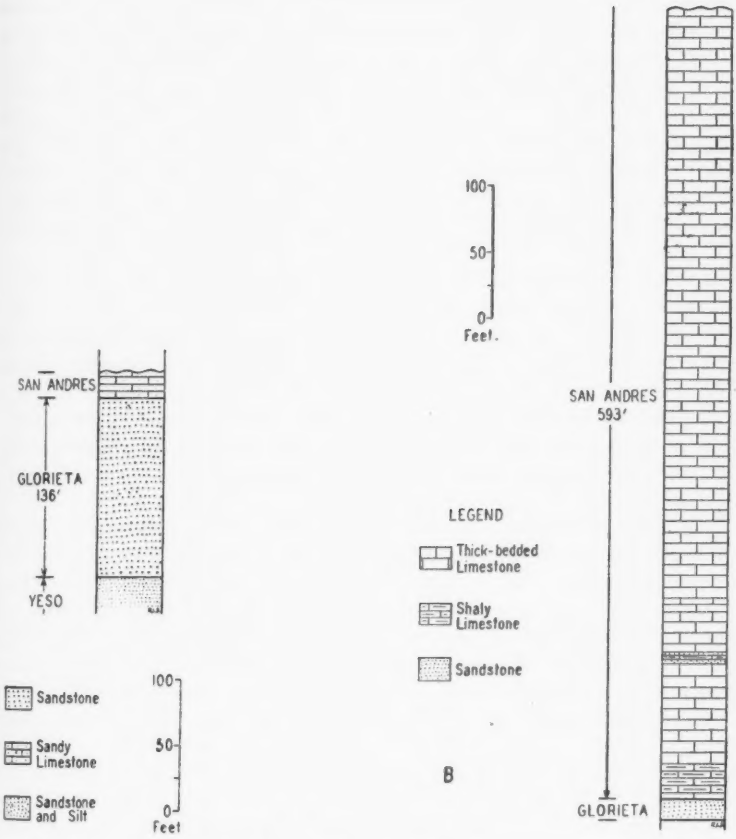

FIGURE 2.—*Columnar sections*

(A) Columnar section of the Glorieta formation at the type locality.
(B) Columnar section of the Glorieta and San Andres formations at the type locality of the San Andres.

ADDITIONAL NOTES

LITHOLOGY: The light color and siliceous character of the Glorieta are uniform and persistent. Cross-bedding is a prominent feature in some exposures—for example, at the Lomo de las Cañas east of Socorro—but at most places the Glorieta does not appear to have been laid down under especially turbulent conditions. The writers feel that a detailed sedimentary study of the formation is desirable. Information on size and shape of the quartz grains, character and distribution of heavy minerals, and significance of the cross-bedding are needed for understanding the origin of the Glorieta.

DISTRIBUTION AND THICKNESS: The Glorieta is essentially coextensive with the underlying Yeso. It attains its greatest thickness in northern New Mexico. Near Willard, Torrance County, it is 278 feet thick; in Mora Canyon, Mora County, 225 feet; in the Zuni Mountains, 287 feet. In the La Joyita Hills north of Socorro the thickness is 102 feet; near the Mesa del Yeso, 75 feet; and in southern Chupadera Mesa, 69 feet. In Rhodes Canyon, in the San Andres Mountains to the south, only 15 feet is present; and near Bent, in the northern Sacramentos, 12 feet. An increase in thickness farther eastward is shown by the presence of 43 feet near Picacho in the Hondo Valley, and 70 feet in the Dunken Dome well in southwestern Chaves County.

CONTACTS: The Glorieta is probably conformable on the Yeso formation. Its contact with the overlying massive San Andres limestone in most places shows no evidence of erosion. In the La Joyita Hills, however, the sandstone is separated from the limestone by a 3-inch bed of pink shale that curves down into cracks in the sandstone. The shale may have been squeezed into the cracks by deformation, or, more likely, emplaced by deposition after solidification and cracking of the sandstone.

PHYSIOGRAPHIC EXPRESSION: The quartzitic character of the Glorieta makes it highly resistant. It caps much of Glorieta Mesa and forms bold escarpments at many localities.

STRATIGRAPHIC RANK: On account of its wide distribution, persistence of lithology, bold topographic expression, and stratigraphic importance, the Glorieta is considered to be a formation.

CORRELATION: According to Charles B. Read (personal communication, 1942) the Glorieta is equivalent to the upper part of the DeChelly sandstone of northeastern Arizona. A number of geologists believe that it is represented in central and north Texas by the San Angelo sandstone. The writers consider the Glorieta identical with the Hondo sandstone of Lang (1937, p. 850). The formation is thought to belong in the Leonard series.

SAN ANDRES FORMATION

ORIGINAL DESCRIPTION

The San Andres limestone was named by Lee (1909, p. 12, 29), from the San Andres Mountains in southern Socorro County. The locality was given as "in the canyon through which the road passes from Engle to Rhodes's ranch." No detailed section was included.

REDESCRIPTION

Through inquiry among the older residents of the town of Engle and vicinity, it has been established that the canyon referred to by Lee is the one now known as Rhodes Canyon. Consequently, the following section was measured there, in the south wall of the canyon 2 miles west of the road fork at the top of Rhodes Pass and 14.6 miles by road east of the Continental Airlines beacon near Engle. The small tributary canyon in which the section was measured extends from near the middle of the north line of sec. 29, T. 12 S., R. 2 E., southward toward the middle of the section. (*See also* Figure 2B.)

Interval	Description	Thickness (in feet)
	San Andres formation (Upper part missing through erosion)	
10.	Limestone, dark gray to nearly black, thick-bedded, petroliferous; carries numerous fossils (*Dictyoclostus et al.*), especially in a zone 160 feet above the base; contains many small cavities lined with calcite crystals; caps the west slope of the San Andres Mountains	446.0
9.	Limestone, thin-bedded, partly covered	10.0
8.	Limestone, gray to dark gray, dense, thick-bedded, fossiliferous	30.0
7.	Sandstone, gray, argillaceous	2.0
6.	Limestone, gray to brown, dense, argillaceous	2.0
5.	Sandstone, light-colored, coarse, massive	2.5
4.	Limestone, gray, medium- to heavy-bedded, fossiliferous	75.0
3.	Limestone, gray, thin-bedded; partly covered	22.0
2.	Limestone, gray, dense, medium-bedded	4.0
	Total thickness of San Andres formation	593.5
	Glorieta sandstone	
1.	Sandstone, light yellow, medium-bedded	15.0

ADDITIONAL NOTES

Lithology: The San Andres in general is remarkably uniform in lithology over central and southeastern New Mexico. It is chiefly calcitic limestone, with some dolomitic beds. Cherty layers are common in the Guadalupe Mountains region, and much chert is present on the formation's upper surface in the Zuni Mountains. At a few places, as along the road between Fort Stanton and Bonnell's Ranch in southern Lincoln County, considerable pore space is evident in the limestone. In the subsurface of the northeastern part of the State the San Andres changes laterally into alternating limestones, gypsums, and redbeds.

Distribution and Thickness: In the north part of the State the San Andres is thin and crops out discontinuously. It is 110 feet thick at Bluewater Canyon in the Zuni Mountains, and only 15 feet thick in Glorieta Mesa (Fig. 2A). Toward the south and east, however, the formation thickens and assumes a remarkable continuity of outcrop. It is at the surface over more than 5000 square miles in central and southeastern New Mexico, where it caps such regional features as Chupadera Mesa, the Sacramento Mountains, and the Guadalupe Mountains.

Although the San Andres is so widespread, it is difficult to find a place where its full thickness is measurable. Probably such a section exists in the northern Guadalupe Mountains, but field work is insufficient there to furnish the information. Elsewhere the relatively weak redbeds and gypsums of the overlying Whitehorse group have been forced by erosion to retreat far down the dip slope of the San Andres, so that it is 50 miles or more from the base of the formation on the west to the place where it passes under younger beds in the Pecos Valley to the east. The problem is complicated in the valley bottom by the presence of alluvium that masks the upper boundary and by the fact that considerable erosion took place on the San Andres both in pre-Whitehorse time and in the Quaternary. East of the Pecos, a number of wells have been drilled through the San Andres, and they furnish data on its thickness. On the basis of wells drilled near Roswell, Nye (1933, p. 87) states that the thickness of the San Andres ranges from 485 to 750 feet where it is covered by alluvial deposits, and from 813 to 1200 feet where it is overlain by the Whitehorse group. Twelve hundred feet is an average figure for the west side of the Permian Basin.

CONTACTS: Except for the exposure in the La Joyita Hills mentioned under the discussion of the Glorieta, the lower contact of the San Andres appears to be conformable. Its upper surface, however, is an unconformity of some magnitude. In southern Lea County a thin sandstone, the Lovington member of the San Andres (Hills, 1942, p. 270–271), is found at widely varying distances below the top of the formation, indicating post-San Andres erosion and possibly the development of sinkhole topography. In the deep parts of the Permian Basin the San Andres is overlain by lower Whitehorse beds; along the northwest edge of the basin, by upper Whitehorse beds; and still farther northwest, by beds of Triassic age. In the Zuni Mountains, Triassic beds lie on an uneven San Andres surface which is heavily studded with chert and appears to be a typical silicified erosion surface. That the San Andres may have been entirely removed in places by pre-Triassic erosion is indicated by the record of a well recently drilled on the Ordnance Depot grounds at Wingate, McKinley County. According to C. V. Theis (personal communication, 1942) Chinle shale of Triassic age was underlain directly by Glorieta sandstone, thus showing complete absence of San Andres at the northwest end of the Zuni Mountains.

SANDSTONES NEAR BASE: At many places the writers have found one or two thin sandstones in the lower part of the San Andres formation. At the mouth of Bluewater Canyon in the Zuni Mountains the sandstones are 1.5 and 12 feet thick and lie 15 feet and 31.5 feet, respectively, above the base of the San Andres. On Chupadera Mesa one sandstone was found which is 15 feet thick and is separated from the Glorieta by 20 feet of typical San Andres limestone. Thickness and position of the sandstone beds at the type locality of the Yeso and in Rhodes Canyon are shown in Figures 1B and 2B.

In most places these sandstones are like the Glorieta. However, the limestone that separates them from the Glorieta is of the thick-bedded gray dense San Andres type, and the sandstones are thought to belong with the San Andres. Probably they record intervals in early San Andres time during which the sedimentary environment returned briefly to a Glorieta-like phase.

CORRELATION: The correlation of the San Andres is at present under vigorous debate. It is thought that solution of the problem will result from detailed field work in the central Guadalupe Mountains, where facies changes in the San Andres are well exposed. The formation was termed the Picacho limestone by Nye (1933, p. 55), but this name is now abandoned. The San Andres is believed by the writers to belong in the upper part of the Leonard series and thus to be equivalent to the upper part of the Bone Spring limestone of the Delaware basin.

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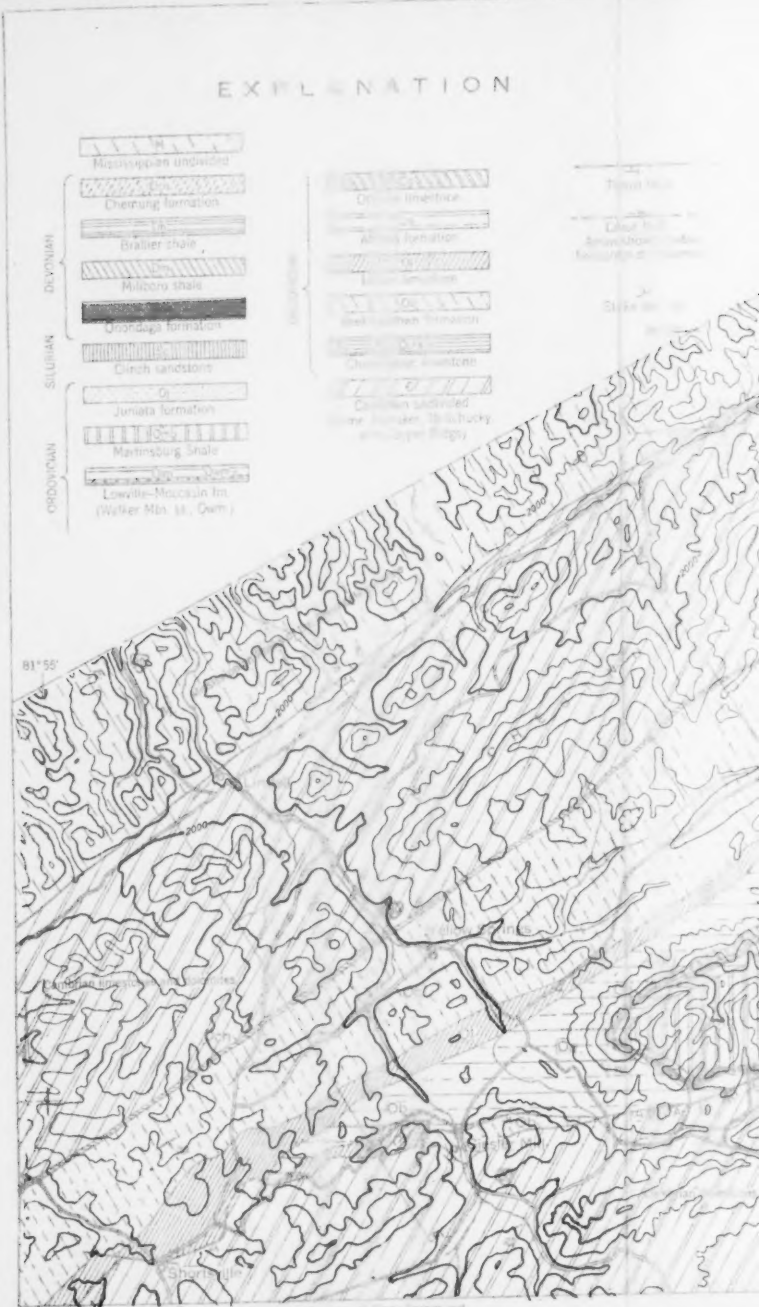

Base from U. S. Tennessee Valley Authority topographic maps of Cooke Springs and Hasters Gap quadrangles and Chilhowie photomosaic index

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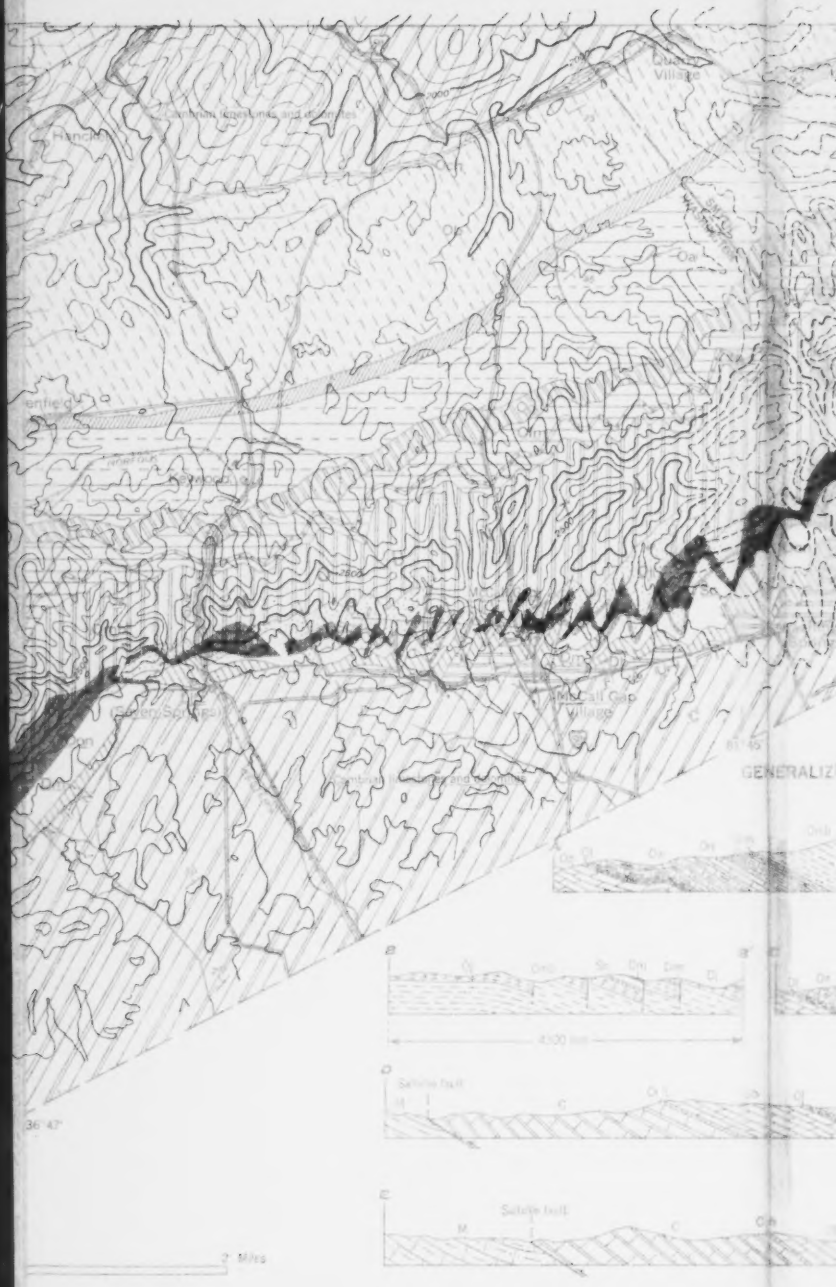
Southwestern end of Walker Mountain, Virginia



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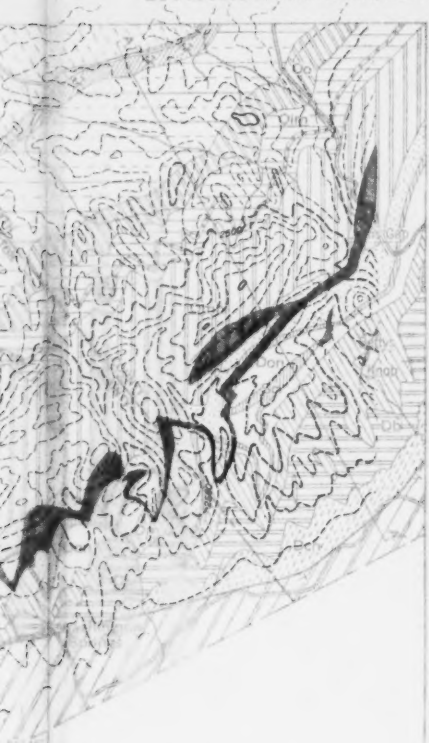

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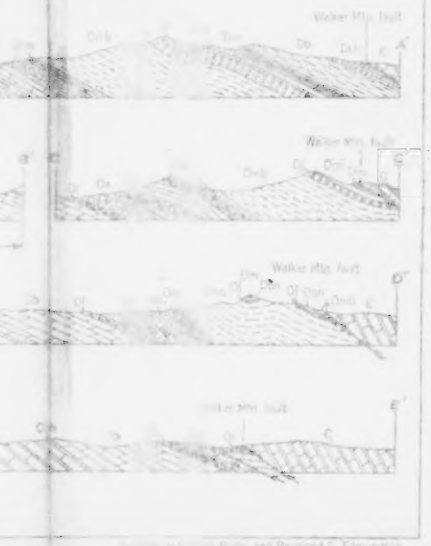

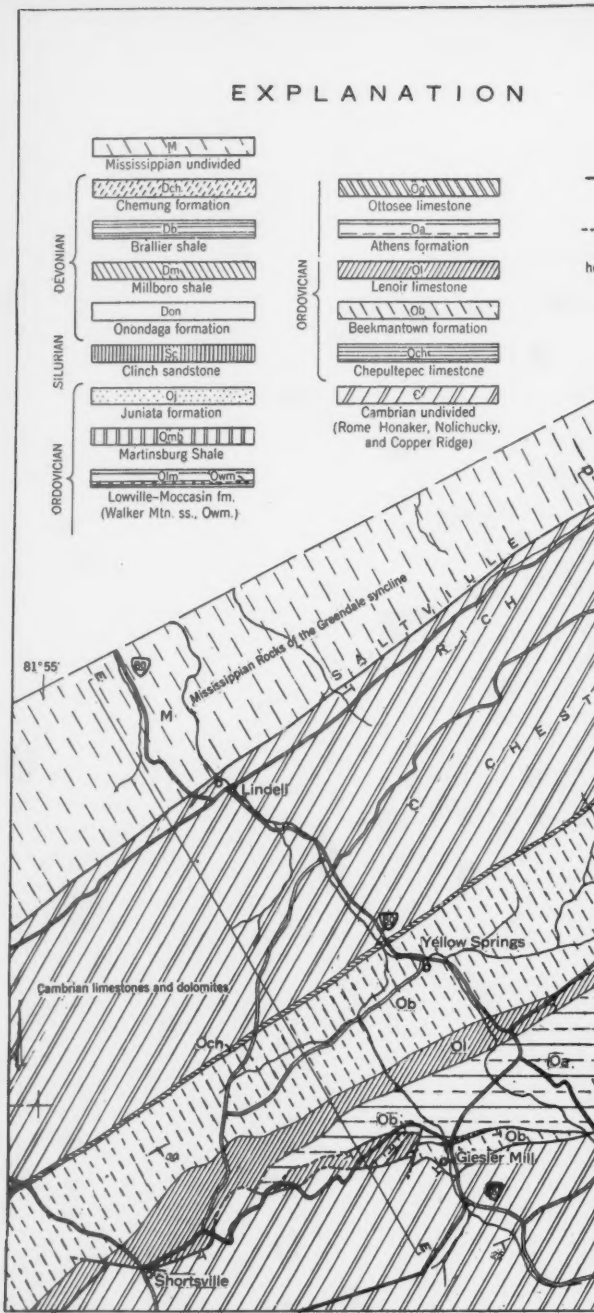

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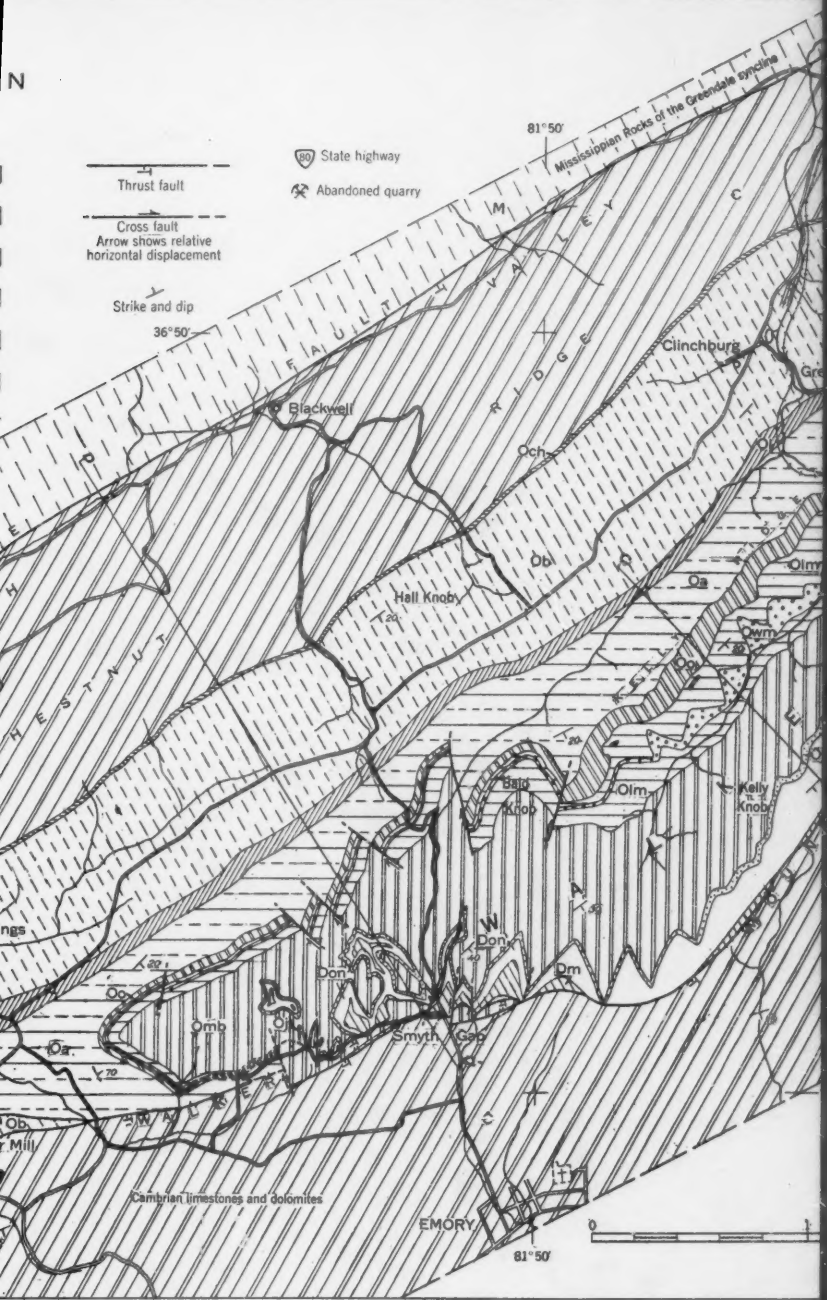

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GEOLOGIC MAP AND SECTIONS
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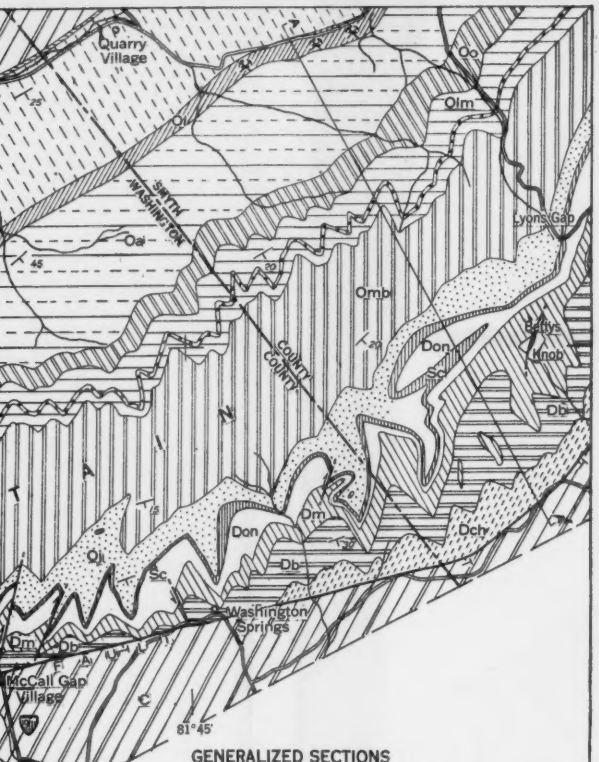

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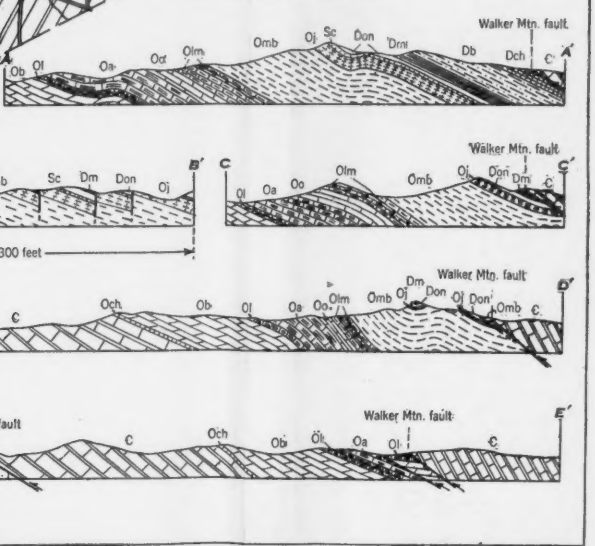

Geology by Charles Butts, and Raymond S. Edmundson

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GEOLOGY OF THE SOUTHWESTERN END OF WALKER MOUNTAIN, VIRGINIA

BY CHARLES BUTTS AND RAYMOND S. EDMUNDSON

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ABSTRACT

Walker Mountain, in Washington County and adjacent parts of Smyth County, Virginia, is on the northwest limb of a major syncline which is bounded on the northwest and southeast by the Saltville and Walker Mountain overthrusts, respectively. The formations, Middle Cambrian to Upper Devonian, composing the fault block have a maximum aggregate thickness of about 9500 feet. The most significant geologic feature near the southwestern end of the ridge is a great unconformity marked by the absence of all the Silurian and a part of the Lower Devonian strata.

Southwest along the Walker Mountain fault, the overthrust Cambrian dolomite is in contact with successively older formations in the upper part of the Walker Mountain block. Near Lyons Gap, the overthrust Cambrian dolomite lies upon Upper Devonian shales and near the southwestern end of Walker Mountain upon the Lower Ordovician limestones.

The regional structure is complicated by small folds with axes parallel to the strike of the beds in the fault block; by cross folds which trend northwest-southeast; and by cross faults, some of which cut both overridden and overthrust rocks and locally offset the trace of the Walker Mountain fault. It is thought that the cross faults represent shears produced during the formation of the cross folds by local rotational stresses in the horizontal plane. Perhaps the cross stresses were developed in the Walker Mountain block during the waning advance of the overthrust mass along an irregular surface of the Walker Mountain fault.

INTRODUCTION

The Appalachian Valley and Ridge province, which includes the Walker Mountain area, lies between the northwestern foot of the Blue Ridge and the eastern escarpment of the Appalachian plateaus. Two divisions have been distinguished—the Great Valley on the southeast and a wide belt with linear parallel ridges, separated by narrow valleys, called the Valley Ridges, on the northwest. In southwestern Virginia the Abingdon-Dublin plain is coextensive with the Great Valley and Walker Mountain, in places flanked on the east by such subordinate ridges as Cloyds, Little Walker, and Brush mountains, is the southeasternmost belt of the Valley Ridges.

Walker Mountain is a narrow ridge that extends about 110 miles from the vicinity of Giesler Mill, Washington County, northeastward to Newcastle, Craig County. From New River nearly to Newport the ridge is named Gap Mountain, and from Newport northeastward it is Sinking Creek Mountain. For most of this distance the crest is between 3000 and 3500 feet above sea level and from 1000 to 1500 feet above the adjacent valleys. A few conspicuous knobs rise above an altitude of 3500 feet. The highest point along the Mountain is in the "Big Bend," about 6½ miles southwest of the village of Bland, Bland County, at an altitude of 4000 feet and about 1500 feet above Stony Fork of Reed Creek.

Northeastward from McCall Gap, Washington County, the crest of the mountain coincides with the outcrop of the Clinch sandstone, whose resistant beds, dipping steeply to the southeast, may be said to be the immediate cause of the mountain.

Walker Mountain is the northwestern limb of a large syncline, the southeastern limb of which has been thrust northwestward and removed by erosion. This overthrust has been named Walker Mountain fault (Stephenson, 1881, p. 227) on account of its development along the southeast base of Walker Mountain in southwestern Virginia. The northwestern limb of the syncline is terminated by a major overthrust—the Saltville fault (Stephenson, 1885, p. 121) which is comparable in extent and displacement to the Walker Mountain fault. The trace of this fault lies in Rich

FIGURE 1.—*Index map showing extent of Walker Mountain and area (ruled diagonally) described*

Valley, a mile or so northwest of Walker Mountain, and extends the entire length of the ridge. Along this fault, wide belts of Cambrian limestone and dolomite, representing the lower part of the Walker Mountain block immediately east of the Saltville fault, are thrust northwesterly upon Mississippian formations of the Greendale syncline. Southeast from the Saltville fault the sequence of the formations is normal, and the dip is southeastward as far as the Walker Mountain fault. These two overthrusts define a fault block that is dominated topographically by Walker Mountain. Among the most important geologic features in the area is a great unconformity with the Silurian and part of the Devonian strata absent. This marked unconformity and the faults were recognized by G. W. Stose and were shown by him on a manuscript map.

Immediately southwest of Walker Mountain (Pl. 1), but along the same general strike, the rocks are sliced by faults, and the major structure is largely obscured. The syncline reappears, with Ordovician formations on its southeast limb intact, a few miles south of Kingsport, Tennessee, and persists southwestward as Bays Mountain to Bulls Gap, Tennessee.

The mountain viewed from a distance appears as a straight even-crested ridge. Actually, the crest is sinuous and characterized by many low knobs. This configuration results largely from the headward erosion of tributaries to North and Middle forks of Holston River. The slopes of Walker Mountain are deeply trenched by ravines separated by narrow spurs. The ravines offset the summit of the mountain from place to place and thus give rise to the sinuous crest. Some transverse streams have cut deep gaps (Pl. 2, fig. 1) through the mountain which afford favorable locations for highways. The slopes above these gaps as well as some of those on the northwest side of the mountain, eroded on the Martinsburg shale, are very steep.

FIELD WORK AND ACKNOWLEDGMENTS

The present paper is based on 2 months of field work, using Tennessee Valley Authority 7½-minute topographical sheets. Additional mapping on the surrounding Abingdon 30-minute sheet, which contributed in part to the geologic interpretations of the southwestern end of Walker Mountain, is based on 4 seasons of field study by the senior writer, 2 by the junior writer, and additional work by geologic assistants. Mr. G. W. Stose of the U. S. Geological Survey made available his manuscript maps covering a part of the Walker Mountain area and offered suggestions for revision of the manuscript. The writers are also indebted to Professor C. M. Nevin of Cornell University for reading the section on structure and to various members of the Virginia Geological Survey for helpful criticism. The discussion of structure and the preparation of the maps and sections are by Edmundson; the stratigraphic interpretations and correlations are by Butts.

STRATIGRAPHY

GENERAL STATEMENT

The formations of the Walker Mountain area are listed in Table 1. All the Ordovician formations down through the Athens limestone crop out on the summit, the northwest slope, and the south end of Walker Mountain; the Silurian and Devonian

formations are restricted to the southeast side of the ridge, adjacent to or close to the Walker Mountain fault. The rock units below the Athens occupy the valley and rolling country immediately northwest of Walker Mountain and extend to the Saltville fault. Pre-Chepultepec rocks, including Honaker, Nolichucky, and Copper

TABLE 1.—*Formations of the Walker Mountain area in southwestern Virginia*

Period and system	Formation
Devonian	Chemung formation Brallier shale Millboro shale Onondaga formation Oriskany sandstone
Silurian	Clinch sandstone
Ordovician	Juniata formation Martinsburg formation Lowville-Moccasin formation Ottosee limestone Athens limestone Lenoir limestone Mosheim limestone Murfreesboro (?) limestone Beekmantown formation Chepultepec limestone
Cambrian	Undifferentiated limestones and dolomites; Copper Ridge dolomite at top

Ridge formations, are grouped and mapped (Pl. 1) as Cambrian limestones and dolomites since they do not crop out on Walker Mountain.

The Copper Ridge dolomite, uppermost Cambrian, delineates the base of the Ordovician system at localities where the thin Chepultepec limestone (lowermost Ordovician) does not crop out.

CAMBRIAN SYSTEM

Copper Ridge dolomite.—The Copper Ridge is almost entirely dolomite northwest of the Saltville fault; to the southeast, including the Walker Mountain area, a few beds of limestone are present. The appearance of limestone in the belt that crosses the Saltville branch of the Norfolk and Western Railway just south of Hanckel (Pl. 1) and extends the full length of the area mapped may reasonably be interpreted as the beginning of the transition from the Copper Ridge dolomite facies to the Conococheague limestone facies of this stratigraphic unit. The Copper Ridge dolomite in this area is about 1400 feet thick. Beds of coarse-grained, friable sandstone, from a few inches to a few feet thick, characterize the Copper Ridge dolomite and the Conococheague limestone and aid in correlation.

ORDOVICIAN SYSTEM

Chepultepec limestone.[1]—The Chepultepec consists of thin beds of limestone and dolomite aggregating about 50 feet. It overlies the Copper Ridge dolomite and underlies limestone and dolomite of the Beekmantown formation. Commonly, a bed of sandstone in the Copper Ridge closely underlies the Chepultepec. The general character and relations of the Chepultepec are shown in Section 1:

Section 1.—At intersection of State Route 91 and Quarry village road, ¾ mile southwest of Quarry Village, Washington County, Virginia

	Thickness Ft.	In.
Beekmantown formation		
7. Dolomite mainly, several hundred feet		
Chepultepec limestone		
6. Limestone with fossils	20	0
5. Dolomite	4	6
4. Limestone, contains *Dakeoceras, Conocerina, Hemithecella, Sinuopea?*	6	6
3. Shale	0	8
Copper Ridge dolomite		
2. Sandstone or sandy dolomite	3	9
1. Dolomite mainly, several hundred feet		

The Chepultepec with its diagnostic curved cephalopods (Bed 4) is exposed on Keywood Branch, 1½ miles northwest of Keywood; on the county road, one-third of a mile southwest of Hall Knob; near Maiden Falls School, half a mile northwest of Shortsville; and at Carvasso Church, along U. S. Route 19, 1 mile southeast of Greendale. These localities are on the belt that parallels Walker Mountain near the northwestern boundary of the area of this paper.

The Chepultepec marks the boundary between the Beekmantown and the Copper Ridge formations along the northwestern belts of the Valley and between the Beekmantown and the Conococheague limestone along the southeastern belts. The outcrop of the Chepultepec (Pl. 1), as well as other formations mapped as narrow belts, parallels closely the trend of Walker Mountain.

Beekmantown formation.—The Beekmantown is almost entirely dolomite in the southern part of the Valley in Virginia. A few thin scattered beds of limestone make but a small proportion of the formation. The width of the outcrop of the Beekmantown in this area is from 3000 to 5000 feet, and its thickness is about 1250 feet. The formation is well exposed along State Route 91 south of its junction with the Quarry Village road and along the road leading northwest from Keywood. The best exposure of the upper half or more is along Stonemill Creek between Clinchburg and Greenfield.

The Beekmantown formation, though not richly fossiliferous, has enough diagnostic genera for wide correlation. The fossils occur in a lower and upper zone. The lower zone, apparently comprising more than the lower half of the formation in southwest Virginia, is marked by such diagnostic genera as *Lecanospira* and *Roubidouxia;* the upper zone is characterized by *Orospira*, *Coelocaulus*, and *Ceratopea*. *Ceratopea* has not been noted in the Walker Mountain area but has been collected from several belts to the northeast.

[1] The Chepultepec is regarded by many geologists as a part of the Beekmantown group.

Murfreesboro (?) limestone.—Along State Route 80 and Smyth Branch about half a mile southeast of Yellow Springs, the Beekmantown dolomite is succeeded by a pink, mottled limestone, about 50 feet thick. This limestone contains small water-worn pebbles of quartz and gray chert and larger greenish pebbles suggesting quartzite. Most of the larger pebbles do not exceed half an inch in diameter. There are also many well-rounded grains of quartz about a millimeter in diameter. These beds are evidently basal to the Mosheim limestone and may partly fill a hiatus that occurs elsewhere in this area at the top of the dolomite. This rock has a very small outcrop and was not separately mapped.

Fossils include abundant small rhynchonellid brachiopods and a few specimens of a bryozoan, regarded as a species of *Nicholsonella*. The presence of any bryozoans evidently shows the post-Beekmantown age of these beds. The beds are tentatively correlated with the basal chert conglomerate of the Blackford facies (Butts, 1940, p. 126, Pl. 23A, B) of the Murfreesboro formation which occurs along Clinch Valley and southwest to the vicinity of Clinchport, Scott County.

Mosheim limestone.—A thick-bedded, compact limestone of Mosheim aspect succeeds the Murfreesboro (?) along State Route 80. A similar high-grade limestone about 20 feet thick crops out along the road to Smyth Gap about 500 feet south of its intersection with the road to Clinchburg. Here the limestone follows the Beekmantown dolomite. No limestone referable to the Mosheim was noted between the locality mentioned above and the Porterfield quarry, a distance of about 10 miles along the strike and several miles northeast of the area shown on Plate 1. At several places northeast of the Porterfield quarry the Mosheim limestone is immediately above the Beekmantown dolomite. At some of these places it is a thick-bedded, compact, pure limestone, 50 feet thick.

Lenoir limestone.—The Lenoir is a blue, medium- to thick-bedded limestone, shelly and nodular in the upper part. It is free of chert nodules which are rather abundant in other areas. Some layers are a mass of ellipsoidal lumps or nodules bound together by a shaly or clayey material. Such a layer at the top of the formation extends for three-fourths of a mile or more northeast of the road going south to Smyth Gap and 500 feet south of the road to Clinchburg. The limestone is largely fragments of fossils, especially bryozoans. The Lenoir crops out in a narrow strip, 300 to 400 feet wide, and is so widely exposed that it can be mapped easily. Since the dips average about 25° SE., the thickness must be 100 to 150 feet.

The Lenoir is rather pure limestone. It has been extensively quarried at Old Quarry Village for use in the Mathieson Alkali Works at Saltville. The quarry, nearly a mile long, probably was abandoned because of the discovery of a large body of purer Holston limestone at Porterfield, 5 miles east of Saltville, from which the limestone for the Alkali Works is now obtained. The most significant fossil of the Lenoir is *Maclurites magnus*, specimens of which were seen all along the outcrop of the limestone in this area. Bryozoans are abundant, but only a few generic identifications have been made. A large massive *Monotrypa*, a *Nicholsonella*, and a form comparable to *Nemataxis* have been provisionally identified. Brachiopods are fairly plentiful in some layers. Species of *Sowerbyella*, *Valcourea*, and *Sowerbyites* have been collected from the Porterfield quarry and identified by G. A. Cooper of the U. S. National

Museum. At the same locality there are also many large specimens of *Endoceras* in the top of the Lenoir, just below the Holston limestone. Many specimens of the same fossil occur in the top of the Lenoir at one place in the old quarry at Quarry Village, where they are immediately overlain by the Athens limestone. This *Endoceras* can be compared with *E. magister* Ruedemann or *E. hudsoni* Ruedemann from the Chazy limestone of the Lake Champlain region in New York. A single specimen of *Spyroceras*, of a different species from *S. clintoni* Miller, was found. *Spyroceras clintoni* also occurs in the Chazy. *Maclurites magnus* affords grounds for correlating the limestone bearing it with the Lenoir limestone of Tennessee, and it appears to be the same as the *Maclurites* so abundant in the Crown Point limestone on Lake Champlain. The fossils, particularly the *Maclurites*, justify the correlation of the Lenoir with part, at least, of the typical Chazy.

Athens formation.—The Lenoir is overlain by the Athens formation, since the Holston and Whitesburg limestones are absent in this area. In Rich Valley (Pl. 1) it is predominantly an evenly thin-bedded, somewhat argillaceous, dark to black limestone with, near the middle, layers of black, compact limestone. This limestone crops out along a secondary road about 1000 feet south of the west end of the old quarry south of Quarry Village. Locally, as shown in an exposure on State Route 80 just north of Giesler Mill, the rock is a fissile, rusty shale at the bottom. The peculiar gray, flaky chips of the weathered limestone are conspicuous in bare fields and in exposures in road cuts. The thickness of the Athens in this area has not been accurately measured. Along the secondary road just south of Quarry Village, the width of the outcrop of the Athens is about 2000 feet, and the average southeast dip is about 20°; thus the thickness is estimated to be about 700 feet. Farther northeast in Rich Valley, where it is crossed by State Route 16 north of Marion, the thickness is about 350 feet.

A few graptolites have been collected from the Athens in this area. At Giesler Mill (Pl. 1) *Dicranograptus*, *Climacograptus*, and *Didymograptus serratula?* were obtained. Near the northeast end of the old quarry, near Quarry Village, *Didymograptus sagitticaulis* is abundant associated with *Nemagraptus gracilis* and a species of *Dicranograptus*. The trilobites *Robergia major*, *Dionide*, and *Triarthrus* were also obtained here. *Robergia*, *Dionide*, and a few species of graptolites are widely distributed throughout the Appalachian Valley and occur in formations that, on the bases of character and stratigraphic relations, are referable only to the Athens.

Ottosee limestone.—The Ottosee limestone overlies the Athens formation. It is largely fragmental, nodular to platy, argillaceous, and weathers into small, thin, lenticular gray fragments. These features are characteristic of the Ottosee throughout its extent in southwestern Virginia and adjoining parts of Tennessee.

The Ottosee crops out near the northwest base of Walker Mountain and extends northeastward at least to the vicinity of Bland, Bland County. Its thickness along Walker Mountain is about 300 feet. The most abundant fossils of the Ottossee in this area are bryozoans, most of which have not been described. They are said by Ulrich, who has studied many collections, to be in the upper part of what has hitherto been classified as Ottossee. He has recognized this zone as far south as Heiskell, Knox County, Tennessee. Other fossils, generally rare and poorly preserved, are

the brachiopods *Strophomena, Sowerbyella, Oxoplecia,* and a few trilobites of the genera *Illaenus* and *Calliops*. Crinoid stem plates are common, and an occasional specimen of the cystoid *Echinosphaerites* has been found. *Receptaculites,* common northwest of Clinch Mountain, probably occurs but has not been found in this area.

Lowville-Moccasin formation[2].—In southwest Virginia and the adjoining northern part of Tennessee, the Ottosee limestone is succeeded by the Lowville-Moccasin formation. The hyphenated name is used because the formation contains both limestone of the Lowville type and red beds characteristic of the Moccasin. The red beds are regarded of Lowville age because locally they include fossiliferous limestone of Lowville type. The lower 50 to 100 feet of the formation in this area is a predominantly blue limestone with layers mottled with red. Northwest of Clinch Mountain a few feet of red beds commonly separate this blue limestone from the underlying nodular Ottosee. The sequence is given in the following section:

Section 2.—Along U. S. Route 19 in Little Moccasin Gap, Russell County, Virginia

	Thickness Feet
Lowville-Moccasin formation	
4. Limestone and mudrock, mainly red, several hundred feet	
3. Limestone, blue; contains *Cryptophragmus*	50–100
2. Mudrock, red	5
Ottosee limestone	
1. Limestone, nodular and shaly, several hundred feet	

The sequence is the same in the Walker Mountain area, except that Bed 2 is absent or covered. On State Route 16, about 6 miles north of Marion and 12 miles northeast of Lyons Gap, the basal limestone (Bed 3) is 50 feet thick and slightly reddish. The part of the Lowville-Moccasin formation overlying the limestone described above is composed chiefly of distinctive red argillaceous-calcareous beds. These beds constitute the Moccasin formation and were named from Big Moccasin Creek east of Gate City, Scott County. On weathered outcrops the Moccasin is shale or shaly mudrock. In fresh exposures, on steep bluffs or in deep road cuts, the rock is limy, as on State Route 91, about three-fourths of a mile north of McCall Gap, Washington County. The alternating red and yellow shaly beds are exposed along the road half a mile north of Seven Springs. Of special significance is the occurrence of two beds of sandstone, separated by about 50 feet of red rock, in the red (Moccasin) part of the formation.

The basal blue limestone member of the Lowville-Moccasin is sparingly fossiliferous, and those fossils that occur seem to be confined chiefly to a few thin layers. *Zygospira recurvirostris* is most common. A slab collected from a high spur a mile southwest of Quarry Village contained *Zygospira recurvirostris* and species of *Doleroides, Calliops,* and *Leperditia*. *Cryptophragmus antiquatus,* the most significant fossil for the correlation of the Lowville, was collected at the north end of Kelly Ridge, about half a mile south of Greenfield, and along the highway 3000 feet north of Lyons

[2] The nomenclature and correlation is based upon observations by Butts made during 35 years of field work along the Appalachian Valley from central Pennsylvania to northern Alabama. Other geologists do not agree fully with him, and the subject has become a controversial one which cannot be appropriately discussed here. The basis of the author's interpretation has been discussed (1940, p. 179–191).

Gap where it is quite plentiful. *Tetradium cellulosum*, regarded as diagnostic of the typical Lowville limestone of New York (formerly known as the Birdseye limestone), and *Cryptophragmus* are in close association the Appalachian Valley generally, but *Tetradium* has not been found in this area.

WALKER MOUNTAIN SANDSTONE MEMBER: The name Walker Mountain sandstone is here proposed for the upper of the two sandstones in the Moccasion formation. It is a thick-bedded, gray, quartzose rock 8 to 12 feet thick (Pl. 3, fig. 2). It persists the entire length of the area mapped, swings around the south end of Walker Mountain, and extends northward along the southeast slope for about half a mile to where its outcrop is cut off by a transverse fault. The Walker Mountain sandstone is an important map unit and locally makes conspicuous topographic features. It is best displayed along Keywood Branch road (Pl. 3, fig. 2) about half a mile north of Seven Springs and along State Route 91 about a mile northwest of McCall Gap. Other good exposures are on the crest of the mountain for about half a mile north of the point where it is crossed by the road north from Smyth Gap. Thin sandstones near the horizon of the Walker Mountain sandstone member persist northward along the same belt of outcrop at least to a point about a mile north of Newport, Giles County. The Walker Mountain sandstone rises above the surface on the northward-pitching axis of a syncline about a mile northeast of Giesler Mill, Washington County. It has been eroded away on the southwest as far as a point in the Bays Mountain syncline a few miles southwest of Kingsport, Tennessee.[3]

Martinsburg formation.—Shale of the Martinsburg formation, immediately overlying the Lowville-Moccasin formation, is tawny-colored on weathered exposures with intercalations of highly fossiliferous limestone. Much of the formation that looks like shale is a fine-grained, brownish sandstone. The thin layers of limestone are generally a compact mass of brachiopod shells of the genera *Resserella* (*Dalmanella*) and *Sowerbyella* which were probably swept together on the shallow sea bottom by gentle currents.

The outcrop of Martinsburg, occupying the crest and upper slopes of Walker Mountain near the southwestern end, averages about half a mile in width. This width is controlled by the flatiron-shaped areas of younger rocks, especially of the Juniata formation which overlies the Martinsburg shale and crops out on the southeast spurs of the mountain in conformity with its southeastern dip slopes. The width of outcrop of the Martinsburg decreases toward the south end of the mountain due to the upward rise of the formations on the northward-pitching axis of the syncline. The Clinch sandstone, which is an inconspicuous thin bed along the southeast slope of the mountain in the vicinity of McCall Gap, gradually thickens to the northeast and becomes a prominent ridge maker at Lyons Gap. Northeast from Lyons Gap the mountain coincides with the outcrop of the Clinch sandstone, and the belt of Martinsburg is restricted to the northwest slope of the ridge. East and west of Smyth Gap, through a distance of about a mile, a sliver of Martinsburg shale has been thrust upon Devonian shale and in places upon Onondaga chert,

[3] The Walker Mountain sandstone is believed to be the same as the sandstone which Keith (1905, p. 6) called Clinch in Bays Mountain. Both are overlain by the Martinsburg formation.

FIGURE 1. McCALL GAP IN WALKER MOUNTAIN, WASHINGTON COUNTY
Looking north from a point along State Route 91.

FIGURE 2. McCALL GAP
Stippled line marks position of an 8 inch layer of Oriskany sandstone. Below the Oriskany is 18 feet of Clinch sandstone followed by the Juniata formation. The Oriskany is succeeded above by Onondaga shale, sandstone, and a 10 foot bed of massive chert.

FIGURE 3. SMYTH GAP
About 10 feet of cross bedded gray sandstone interbedded with red shale of Juniata lithology.

FIGURE 1. WALKER MOUNTAIN FAULT ABOUT 100 YARDS WEST OF SEVEN SPRINGS
Stick marks position of fault; Honaker dolomite (left) thrust upon Millboro shale. Looking southwest.

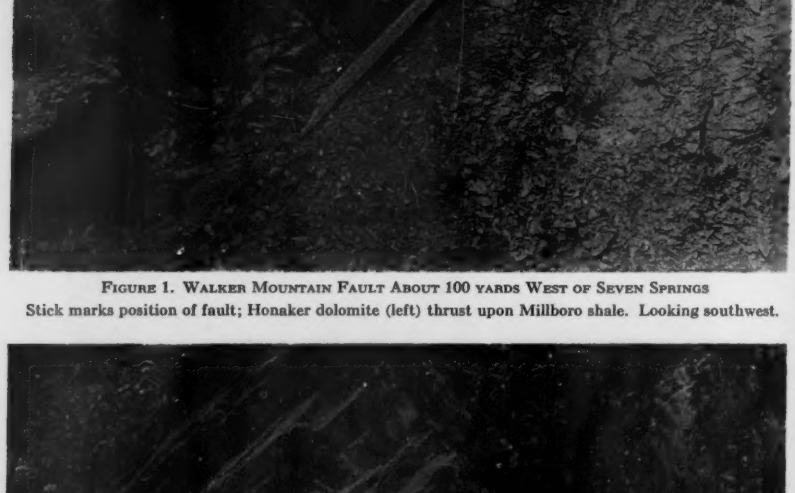

FIGURE 2. WALKER MOUNTAIN SANDSTONE MEMBER OF THE LOWVILLE-MOCCASIN FORMATION ALONG KEYWOOD BRANCH ROAD HALF A MILE WEST OF SEVEN SPRINGS
Looking southwest along the strike.

presumably along a minor branch of the Walker Mountain fault. The identity of this shale is indicated by the occurrence of *Resserella* (*Dalmanella*) and *Sowerbyella* which are rather plentiful in some of the shale at the top of the spur one-fourth of a mile northwest of Smyth Gap. Below this point, on both slopes of the spur, the Onondaga and the overlying black shale crop out. The Martinsburg shale, as measured along State Route 91 north of McCall Gap, is about 1600 feet thick. Minor flexures of the beds cause some uncertainty in determining thickness.

The fauna of the highly fossiliferous limestone layers of the Martinsburg formation in the Walker Mountain belt has not been studied adequately. The few forms mentioned are those recognized casually in the field. The most common and widely distributed species are *Resserella* (*Dalmanella*) *bassleri?* and *Sowerbyella sericea* (Hall). Myriads of valves of these two species are packed together in the thin layers of limestone. *Rafinesquina alternata* and *Zygospira recurvirostris* are also abundant. The fossils cited are common in the Trenton limestone at Trenton Falls, New York, and in the lower part of the Martinsburg in the middle and southeastern belts of the Valley and warrant correlating the lower part of the Martinsburg in those belts with the Trenton limestone. In this area *Cryptolithus tessellatus*, perhaps the most characteristic Trenton fossil, has not been found, but it is present in the lower Martinsburg as far south in Virginia as Buchanan, Botetourt County. In the upper part of the Martinsburg in this area *Orthorhynchula linneyi*, *Hebertella sinuata*, *Rafinesquina alternata*, and *Byssonychia* indicate that the upper beds of the Martinsburg here are of the same age as the Maysville limestone of the Cincinnati region and the upper horizons of the Lorraine shale of New York. In other parts of Virginia the Eden beds, which are between the Trenton and Maysville at Cincinnati, are present. It is assumed that they are also represented in Walker Mountain, but their presence has not been proved. In the northern end of the Valley in Virginia and in central Pennsylvania the Martinsburg is succeeded by the Oswego sandstone which has not been identified in the southern part of the Valley.

Juniata formation.—The Juniata formation in Walker Mountain is composed of brownish-red shale, mudrock, and sandstone. It is about 300 feet thick at Lyons and McCall gaps. Southwestward from McCall Gap it thins to 50 feet at Litz (Severn Springs) and to not more than 10 feet at Smyth Gap. It has not been recognized with certainty at every place within the mountain west of Smyth Gap, although it is mapped as a continuous belt. The Juniata is nonfossiliferous and probably nonmarine.

The Juniata extends northeastward into central Pennsylvania, continues beneath the Allegheny Plateau, and reappears at Queenston at the north end of the gorge of Niagara River where it is called the Queenston shale. Southwestward, as at Cumberland Gap and in Sequatchie Valley, Tennessee, it becomes a red-tinted, fossiliferous limestone and is called the Sequatchie formation. It becomes the highly fossiliferous Richmond formation in southwestern Ohio and southern Indiana.

SILURIAN SYSTEM

Clinch sandstone.—In a complete sequence the Juniata is directly succeeded by the Clinch sandstone. This relationship in Walker Mountain is maintained northeast

of a point midway between Litz and McCall Gap, but southwest of Litz along the mountain, with the possible exception of the vicinity of Smyth Gap, the Clinch and other succeeding formations are absent, and the Juniata is succeeded by the Onondaga formation (Devonian). At Smyth Gap (Pl. 2, fig. 3) about 10 feet of cross-bedded gray sandstone, possibly representing the Clinch, lies between the Onondaga and about 10 feet of red shale that is in the position of the Juniata as it rests directly upon the *Orthorhynchula* zone of the Martinsburg. This sandstone, mapped with the Juniata (?), extends about half a mile southwest of Smyth Gap, where both the sandstone and red shale are apparently cut off by the Walker Mountain fault. In McCall Gap the thin-bedded to medium thick-bedded Clinch sandstone is 18 feet thick, whereas in Lyons Gap, only $3\frac{1}{2}$ miles northeast along the strike, it is 30 feet thick. Below the sandstone at Lyons Gap sandstone and red shale alternate through about 30 feet to the typical Juniata. This suggests that the two formations at this locality grade into each other. About 7 miles northwest of Walker Mountain the Clinch sandstone is 150 to 200 feet thick in Clinch Mountain, along which it forms the backbone of the mountain southwest nearly to the latitude of Knoxville, Tennessee.

The typical Clinch is destitute of fossils except *Arthrophycus allegheniensis*, a supposed worm burrow. At Cumberland Gap, in extreme southwestern Virginia, the Clinch passes into a facies with marine fossils which serve to correlate it with the Albion sandstone of New York and with the Brassfield limestone of Ohio and Indiana.

GREAT UNCONFORMITY AND OVERLAP

Particular attention should be called to the gradual southward thinning and disappearance of the formations between the Martinsburg shale and Onondaga formation. This thinning begins as far north as New River, where the Cayugan and Helderberg formations, except for the Becraft limestone and the overlying Oriskany sandstone, have about disappeared from the section. The thinning continues southwestward and is noticeable south of Wytheville with the thinning of the Clinch and Clinton formations. The Clinton disappears a few miles north of Lyons Gap at the northeastern margin of the area covered by this paper. From Lyons Gap southwestward the Clinch thins in turn and disappears about a mile northeast of Seven Springs. At Seven Springs the Juniata has thinned to about 50 feet, and at Smyth Gap, 3 miles farther southwest, it has thinned to not more than 10 feet. In this part of the mountain the Onondaga is within a few feet of or even in contact with the Martinsburg shale, leaving a stratigraphic gap between the two formations equal to most of the Juniata, the full thickness of the Silurian, and the Lower Devonian Helderberg and Oriskany formations. The interpretation is that, after Martinsburg time, the sea withdrew northward and each younger formation was terminated farther north against the regressive shore line. This shore line trended more to the east than the direction of Walker Mountain as is shown by the following facts: namely, near the south end of Walker Mountain the Onondaga is nearly or quite in contact with the Martinsburg; $2\frac{1}{2}$ miles north of Marion, on a remnant of the southeast limb of the Walker Mountain syncline, the Onondaga is also in contact with the Martinsburg; about 5 miles farther north, near the summit of the mountain, the

Juniata, the Clinch, and the Clinton aggregating several hundred feet are present between the Onondaga and Martinsburg. A line connecting the points of contact of the Onondaga and Martinsburg, representing approximately the old shore line in Onondaga time, trends several degrees farther east than the crest of the mountain.[4] In Onondaga time the region was again submerged, and the marine sediments of the Onondaga were deposited over the entire area in the existing nonsequential relations. A detailed account of this unconformity is beyond the scope of this paper.

DEVONIAN SYSTEM

Oriskany sandstone.—The Oriskany at McCall Gap is a ferruginous, friable, dark-brown sandstone composed of well-rounded quartz grains about a millimeter in diameter. This color and texture distinguish the Oriskany along the southeast slope of Walker Mountain from the vicinity of McCall Gap northeastward to the meridian of Bland, Bland County. The thickness at McCall Gap (Pl. 2, fig. 2) is about 8 inches; scattered boulders between McCall and Lyons gaps indicate a thickness of perhaps 2 feet. The Oriskany has not been noted south of McCall Gap.

At McCall Gap a few fragments of fossils were collected, but only *Spirifer arenosus*, the main guide fossil of the formation, can be specifically identified. A *Spirifer* of the *S. murchisoni* type, a *Meristella* that may be *M. lata*, a *Favosites*, probably *F. shriveri*, and a cyathophylloid coral also occur. North of this area, good specimens of *Rhipidomella musculosa* were collected from this bed. On State Route 16, about 4 miles north of Marion, a bed of Oriskany, perhaps 10 feet thick, immediately overlies about 20 feet at Keefer sandstone in the top of the Clinton. It contains numerous specimens of *Spirifer murchisoni*, *Spirifer arenosus*, and *Rhipidomella musculosa*.

Onondaga formation.—The Onondaga formation in the Walker Mountain area is composed of slightly calcareous sandstone near the base and massive chert in the upper part. This sequence is fairly well established in Section 3.

Section 3.—McCall Gap along State Route 91 about 2 miles north of Glade Springs, Washington County, Virginia

	Thickness Ft.	In.
Onondaga formation		
6. Chert, massive, jagged	10	0
5. Slumped bank with much broken chert, probably thin-bedded chert	35	0
4. Sandstone, thin layers; slightly chertified aggregation of minute quartz grains stained with iron or manganese oxide; contains *Anoplia nucleata, Chonostrophia complanata, Elytha fimbriata, Eodevonaria arcuata, Meristella nasuta?, Orthothetes pandora, Pentagonia* sp., *Pholidops areolata, Spirifer duodenarius, S. planicostatus, S. varicostus, S. segmenta?, Stropheodonta perplana, Platyostoma lineatum*	5	0
Oriskany sandstone		
3. Dark, friable sandstone; a compact aggregation of quartz grains; contains *Spirifer arenosus*	0	8
Clinch sandstone		
2. Sandstone, gray, quartzose, rather thin-bedded	18	0
Juniata formation		
1. Red sandstone and shale		

[4] These relations will be shown on the geologic map of the Abingdon quadrangle, in the course of preparation.

The sandstone (Bed 4) persists southward to Seven Springs where it contains small grains of glauconite and where the following fossils were collected: *Anoplotheca acutiplicata*, *Atrypa reticularis*, *Elytha fimbriata*, *Leptaena rhomboidalis*, *Orthothetes pandora*, *Spirifer divaricatus*, *S. duodenarius*, *S. macrus*, *Stropheodonta perplana*, *Platyceras dumosa*, *Platyostoma lineatum*, *Bollia ungula*, *Favulella* (*Amphissites*?) *favulosa*, and *Octonaria stigmata*.

About 10 feet of these fossiliferous beds is exposed at Seven Springs. The upper and lower contacts of the Onondaga formation are not exposed; however, the upper chert, followed closely by black shale, is present on the adjacent slopes. The fossiliferous sandstone is exposed about 20 feet from the Juniata red shale. It is not known whether Clinch sandstone is present in the covered interval. Section 4 gives the stratigraphic sequence near the south end of the mountain.

Section 4.—Smyth Gap, 3 miles southwest of Seven Springs and 1 mile north of Emory, Washington County, Virginia

	Thickness Feet
Millboro shale	
6. Black shale, interval largely covered	30
Onondaga formation	
5. Chert	20
4. Shale and thin sandstone with streaks of red; no fossils found but resembles the fossiliferous beds (Section 3, bed 4) at McCall Gap	25
Clinch (?) and Juniata (?) formations	
3. Sandstone, cross-bedded, Clinch (?)	10
2. Shale, red, Juniata (?)	
Martinsburg formation	
1. Shale; contains *Orthorhynchula* and *Sowerbyella*	

The only certainly identified parts of this section are the Millboro shale, the Onondaga chert, and the Martinsburg formation. The chert has been traced continuously along the southeast slope of Walker Mountain for many miles to the northeast. It also extends about 1 mile southwest of Smyth Gap to a point on the highest spur of the mountain. Here the exposures are not adequate for precise boundary determinations, but the Onondaga is approximately in contact with the Martinsburg formation. It is evident that at the south end of Walker Mountain most, if not all, of the Juniata, all the Silurian, and the lower Devonian (Helderberg and Oriskany) are absent, probably through nondeposition. The Onondaga is persistent along the southeast slope of Walker Mountain and crops out in a striking flatiron pattern due to the southeast dip of the rocks and the transverse ravines.

The Onondaga age of the formation in this area has been established by its fossils and by tracing (Kindle, 1912) from New York to southern Virginia. Of the fossils cited from the sections at McCall Gap and Seven Springs, *Anoplotheca acutiplicata*, *Eodevonaria arcuata*, *Spirifer duodenarius*, and *S. divaricatus*, as well as *Bollia ungula* and *Favulella* (*Amphissites*?) *favulosa*, occur in the Onondaga formation of New York or Pennsylvania, and they are not reported from any other formation in those States. *Octonaria stigmata* likewise is known elsewhere in the Onondaga limestone at the Falls of the Ohio at Louisville, Kentucky. Nearly all the species cited above were

obtained from a single, small slab of rock at Seven Springs, and members of the same assemblage are closely associated all along the Valley in Virginia.

Millboro shale.—The Onondaga is overlain by a black, fissile shale that represents the Millboro shale (Butts, 1940, p. 308–309) of Marcellus and Naples ages. Not enough is known about this black shale in this area to state its precise stratigraphic equivalence. No fossils have been collected, but in the continuation of its outcrop in Cleghorn Valley northeast of Lyons Gap *Schizobolus truncatus* occurs and probably indicates a Naples age.

The Millboro is present along the lower southeast slope of the mountain. From the vicinity of Smyth Gap northeast to Seven Springs it is in fault contact with the Honaker dolomite. In places for a short distance southwest of Smyth Gap, it is in fault contact with the Martinsburg formation which was brought up and over a part of the Millboro along a subordinate thrust, probably a branch of the Walker Mountain fault. Northeast of Seven Springs and extending beyond the northeastern margin of the area shown on Plate 1, it occurs in normal sequence and is overlain by Brallier shale of Portage age.

The thickness of the Millboro has not been reliably determined, but under the most favorable conditions for measurement, as on the spurs one-fourth to half a mile northeast of Washington Springs, where the Brallier shale is present, the thickness is provisionally determined to be about 300 feet.

Brallier shale.—The Millboro shale is succeeded by the Brallier shale, which grades into the overlying Chemung formation. The Brallier is composed predominantly of thin-bedded to laminated sandstone and siliceous gray shale. At or near the bottom, locally, is a sandstone about 10 feet thick, which makes knobs on the crests of the spurs between McCall and Lyons gaps. Exposures revealing its characteristic features are on State Route 80, one-fourth of a mile south of Hayters Gap; on State Route 42, for a mile immediately west of Allison Gap, Smyth County; and especially in Hungry Mother Park north of Marion, Smyth County. The Brallier in the Walker Mountain area crops out from a point half a mile east of Seven Springs northeastward to the vicinity of Lyons Gap. Its belt of outcrop, like that of the Onondaga, makes a number of flatirons. In the vicinity of McCall Gap the outcrop is affected in a number of places by cross faults (Pl. 1). The thickness of the Brallier cannot be determined exactly, but it is between 500 and 1000 feet.

No fossils have been found in the Brallier in this area, and fossils are scarce everywhere. The few that have been found elsewhere, as farther north in Virginia, are mainly survivals of the preceding Naples fauna and indicate a Portage age. The lithology, stratigraphic relations, and few fossils of the Brallier indicate its equivalence with the Hatch and Gardeau shales of the typical Portage group in New York. Generally speaking, the Brallier also corresponds to the lower half of the Kimberling shale of Campbell (1894, p. 171, 177, Pl. 4) and the lower half of the Jennings shale of Darton (1892, p. 13, 17, 18). The name Brallier shale has been applied to these rocks in the central part of the Appalachian Valley inasmuch as they include less than the original Portage group, less than the Kimberling or Jennings formations, and more than either the Gardeau or Hatch shales.

Chemung formation.—The Brallier grades upward into the Chemung formation.

The Chemung as a body differs from the Brallier in containing numerous fossils and a few layers of quartz conglomerate. Much of the shale of the Chemung is yellowish mudrock, whereas the shale characteristic of the Brallier is commonly subfissile, stiff, and more or less sandy and micaceous. The sandstones of the Chemung are coarser and thicker than those in the Brallier. The two formations can generally be identified with reasonable certainty a short distance either above or below their zone of transition.

The Chemung occupies a fringe along the southeast base of Walker Mountain from the vicinity of Washington Springs northeastward to the east margin of this area. Along this belt it is in fault contact with Cambrian dolomite on the southeast. The formation is nowhere well exposed in this area, but scattered, incomplete exposures along the ravines crossing the belt of outcrop afford some information about its character. A thin conglomerate layer was noted along the Sulphur Spring road a mile west of Sulphur Spring Church. The formation is exposed for a few hundred feet along the hard-surfaced road half a mile south of Lyons Gap. The best exposure in the general region, although not within the area of this report, is on Carlock Creek about 2 miles east of Lyons Gap. Here, several hundred feet of thin-bedded sandstone below and 200 to 300 feet of thick-bedded, fine-grained sandstone with some reddish layers and a few layers of conglomerate above, are well exposed down to the Walker Mountain fault which is 500 feet north of Carlock Church. The thickness of the Chemung on Carlock Creek, west of the Walker Mountain fault, is about 1000 feet.

In the lower part of the fine-grained sandstone on Carlock Creek, apparently not far above the Brallier shale, a thin layer of sandstone contains *Chonetes scitulus*, *Camarotoechia congregata*, and *Leptodesma* cf. *L. potens*. This is a characteristic Chemung assemblage. Farther northeast, on the crest and southeast slope of Brushy Mountain (Fig. 1), the same fossils, together with *Productella lachrymosa*, *Modiola* cf. *M. praecedens*, and other small pelecypods, occur in the same kind of sandstone as on Carlock Creek. These fossils are Chemung and occur in beds which are on the same belt and in strike with those identified as Chemung in the Walker Mountain area southwest of Lyons Gap.

STRUCTURE

GENERAL STATEMENT

The dominant geologic structures of the Appalachian Valley and Ridge province, of which Walker Mountain is a part, are folds trending northeast and parallel thrust faults dipping southeast and extending in many cases more than 100 miles. Where two or more of these faulted folds have been crushed together they have formed fault blocks which, overlapping like shingles on a roof, have been described as shingle blocks. Campbell (1925, p. 31–42) describes Walker Mountain and the subordinate Brush (Fig. 1), or Little Walker, Mountain along part of its eastern flank as the dominant topographic features of a shingle block that developed from an anticlinal fold in Sinking Creek valley in Craig County and extends southwestward for about 100 miles to a point in Washington County where it is concealed by the encroachment

of the Walker Mountain (Pulaski) fault[5] which delimits the Walker Mountain block on the southeast side. In the vicinity of New River (Fig. 1), as between Goodwins Ferry and Churchwood, about 60 miles northeast of the area described in this paper, the stratigraphy and structure are summarized (Campbell, 1925, p. 39) as a block consisting of Shenandoah limestone[6] along the lower or northwestern edge of the block and which is terminated on the upper or southeastern side by the early Mississippian Maccrady formation. From the northwestern limit of the Walker Mountain block (Saltville fault) to Brush Mountain on the southeast the rocks dip consistently at about 55° SE.; farther southeast the dips decrease until the rocks in places are nearly horizontal in the Price Mountain anticline southwest of Blacksburg. Campbell points out also that in this area it may be questioned whether the strata constituting the northwest limb of the Walker Mountain syncline are cut off to the southeast by the thrust fault, whose trace extends along the southeast base of Brush Mountain, or whether the syncline is complete beneath the overthrust mass to the southeast. It is probable that the rocks on the northwest limb of the Price Mountain anticline, about 5 miles southeast of Walker (Gap) Mountain (Fig. 1), are the continuation of the same beds which crop out on the southeast slope of Brush Mountain.

The principal structures of the area discussed in this paper (Pl. 1) are similar to those described for the central and northeastern parts of Walker Mountain. The same problem exists regarding the attitude of the beds and their limits beneath the thrust mass on the southeast. The Walker Mountain block is bounded on the southeast by the Walker Mountain fault and on the northwest by the Saltville fault. The trace of the Saltville fault separates Cambrian (Honaker) dolomite of the fault block from Mississippian rocks in the Greendale syncline on the northwest. Honaker dolomite along and southeast of the Walker Mountain fault immediately south of Lyons Gap is thrust upon Upper Devonian (Chemung). Farther southwest along the mountain (Pl. 1) the fault veers westward and progressively eliminates or conceals older formations from the west limb of the syncline.

The abrupt southwestern termination of Walker Mountain, at a point about a mile northeast of Giesler Mill, is due to the erosion of certain ridge-making formations which rise to the surface and swing around the axis of a syncline pitching to the northeast. The effect of this structure is locally important in delimiting the ridge. The fault trace, from the vicinity of Shortsville (Pl. 1) to the Tennessee line, is northwest of the Walker Mountain axis. Along this belt Honaker dolomite is thrust up... Lower Ordovician formations which, along the strike to the northeast, occupy the broad valley west of Walker Mountain. A few miles south of Kingsport, Tennessee, the Walker Mountain syncline reappears with Ordovician formations intact on the southeast limb and persists southwestward as the syncline of Bays Mountain.

Many small folds, with their axes parallel to the regional structure, and at least three cross folds locally disturb the regularity of the general northeast strike of the

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[5] Since the Walker Mountain fault was named from this part of Walker Mountain and has priority in the literature, it is used in this discussion. It is in the southwestward continuation of the Pulaski zone of faulting, which is disrupted around the Marion dome (Cooper, 1936, p. 162–164). It seems to be an offset branch of the Pulaski fault which extends northeastward to a point near Greenville, Augusta County, Virginia.

[6] Includes Lower Cambrian to Middle Ordovician rocks. Copper Ridge dolomite (Upper Cambrian) crops out east of the fault and limits the base of the block in this section.

rocks that form the northwest limb of the Walker Mountain syncline. Cross faults that offset the strata in the upper part of the Walker Mountain fault block are of special interest. Probably structures similar to these, and to some of the larger structures mentioned above, exist throughout the belts of Cambrian limestones and dolomites of the broad valley east of Walker Mountain, but the lack of key beds and generally poor exposures make precise determination of the structures impracticable.

Since incomplete exposures make precise data impossible, the structure sections (Pl. 1) are generalized as to the number and extent of minor folds, thickness of certain formations, and the attitude of the thrust faults.

FOLDS OF THE WALKER MOUNTAIN BLOCK

Within the Walker Mountain fault block the rocks dip to the southeast, except at a few places along the ridge where minor flexures are superimposed upon the major structure and at the southwestern end of the mountain where Upper Ordovician formations wrap around the axis of a northeastward-plunging syncline. Sections across the block from the vicinity of Lyons Gap southwest to Smyth Gap show variations in dip within a horizontal distance of a few hundred feet. The succession of the beds across the ridge point unmistakably to a synclinal structure, but the degree of dip does not lessen as the synclinal axis on the southeast is approached.

Walker Mountain, with an average height of about 400 feet above the adjacent valleys, ends at a point 1 mile northeast of Giesler Mill. At this locality the syncline, with Martinsburg shale occupying the axial position, is complete. In the absence of the ridge-making Juniata and Clinch sandstones, the crest of the mountain, at the apex of the fold and continuing along the southeast limb to a cross fault about half a mile southwest of Smyth Gap, coincides with the outcrop of the Walker Mountain sandstone member of the Lowville-Moccasin formation. Along the northwest limb of the syncline near the southwestern end of the fold the rocks dip about 20° SE.; the same beds in about the same position on the southeast limb are overturned and dip steeply southeast. Farther east, in proximity to the Walker Mountain fault, the strikes and dips are variable. Several of the formations are severely crushed and thinned.

Minor flexures, with axes trending northeast, are superimposed upon the northwest limb of the Walker Mountain syncline at several localities, but only a few are mentioned. Between the Washington-Smyth county line and Bettys Knob to the northeast, a tongue of Onondaga chert occupies the trough of a minor syncline. It is surrounded, except at its northeastern end, by Clinch sandstone. South of the county line the same formations are involved in a minor syncline that is disarranged by a diagonal fault. Within the large expanse of Martinsburg shale between Lyons and Smyth gaps and throughout the distribution of the formation generally are minor flexures which repeat certain beds, but the lack of key markers makes it difficult to match these structures at intervals along the strike. On the prominent knobs northwest of Smyth Gap are small exposures of black shale and underlying Onondaga chert which do not connect with similar rocks along the southeast base of the ridge. The section (Pl. 1, section D–D′) showing the distribution of these rocks and their

relation to the topography indicates minor flexures. East of Smyth Gap the distribution of the formations and their structure make it necessary to assume that the position of the axial line of the Walker Mountain fold is at least as far south as the trace of the Walker Mountain fault.

A conspicuous recess or re-entrant is developed in the regional strike of the rocks (Pl. 1) along an axis of a gentle cross fold that extends northwest of Seven Springs. Additional gentle cross folds, one with an axis almost parallel to the Smyth-Washington county line on the northeast and another with an axis extending northwest of Bald Knob at the southwest, have produced salients. These broad gently curving folds, plunging southeast, affect the regional strike of all the formations within the fault block.

WALKER MOUNTAIN BOUNDARY FAULTS

Walker Mountain fault.—The Walker Mountain thrust in this area is a continuous fault, with some split faults forming slices, that extends along the southeast side of the Walker Mountain ridge (Pl. 1). Faulted slices occur near Smyth Gap and Giesler Mill (Fig. 2). The trace of the Walker Mountain fault, within limits adequate for detailed mapping, can be followed continuously near the southwestern end of Walker Mountain, but clearly defined exposures of the fault surface are lacking. In the vicinity of Seven Springs (Litz), about 300 feet west of the Saltville Branch of the Norfolk and Western Railway, Honaker dolomite is in contact with Millboro shale (Pl. 3, fig. 1). The fault dips 45° SE., but the small exposure of dolomite may be a slice detached a few feet from the main mass to the southeast. The attitude of the fault surface probably varies widely at different localities along the ridge.

The course of the Walker Mountain fault from the Virginia-Tennessee line to the vicinity of Giesler Mill, Washington County, conforms in general to the structural trends; farther northeast the fault trace veers more easterly and crosses the post-Beekmantown formations of the upper part of the fault block. Northeast of the area described here, the trend of the fault is more irregular, and at several localities it deviates far enough from the base of Walker Mountain to permit the introduction of a subordinate ridge variously known as Brush or Little Walker or Cloyd Mountain.

The Walker Mountain fault in the extreme northeastern part of the area, about half a mile south of Lyons Gap, strikes N. 35° E. From this locality to the vicinity of McCall Gap, the trace of the fault defines a southward-curving broad arc that passes about 500 feet south of Washington Springs and progressively changes its trend to almost west at McCall Gap. This westerly trend of the overthrust gradually conceals the Chemung formation near Washington Springs and the Brallier shale in the vicinity of McCall Gap. The structure sections (Pl. 1) are generalized, since the attitude of the beds beneath the thrust mass east of the fault trace and the inclination and behavior of the fault surface are not known. For example in structure section A–A′ it is assumed that the Chemung formation dips more gently than the fault surface. Such an assumption allows narrowing of width of the Chemung by faulting. If the beds dip more steeply than the fault surface, more of the Chemung, and possibly even younger beds, occurs in the section and is covered by the overthrust mass.

From McCall Gap to a point on a high spur, one-fourth of a mile west of Seven

Springs, the Walker Mountain fault trends west. The trace of the fault is offset by four cross faults. On the knob just west of Litz the trace of the fault changes rather abruptly from a west to a southwest course, which is maintained for about 2 miles to the southwest. From a point a mile east of Smyth Gap to the southwestern boundary of the Walker Mountain area the line of faulting, marked by the Walker Mountain fault, or by this fault and minor branch faults, trends generally S. 75° W.

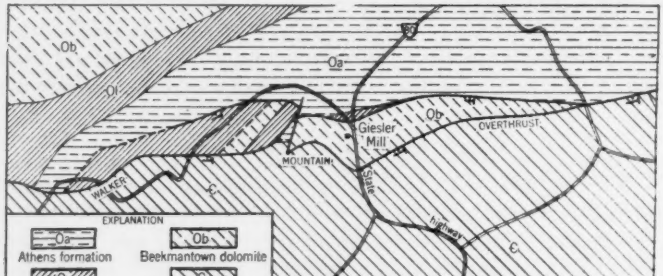

FIGURE 2.—*Sketch of fault slices along Walker Mountain overthrust near Giesler Mill, Washington County, Virginia*

A thin sliver of Martinsburg shale and a few beds of red shale and gray sandstone of Juniata lithology crop out just south of Smyth Gap and extend along the base of the mountain for about a mile. To the north, these beds are thrust, presumably along a branch of the Walker Mountain fault, upon Millboro black shale and in places upon Onondaga chert; to the south they are overridden by Cambrian rocks along the Walker Mountain fault. At a point about half a mile east of Giesler Mill, this major fault splits (Fig. 2) into two branches, each of which involves considerable stratigraphic displacement. These faults coalesce about half a mile east of Shortsville and continue to the southwest as a single fault.

Stratigraphic displacement along the Walker Mountain fault diminishes more or less gradually from a maximum of about 8500 feet at a point south of Lyons Gap to not more than 2000 feet in the vicinity of Shortsville.

Saltville fault.—The northwest boundary of the Walker Mountain block is the Saltville fault which in extent and stratigraphic displacement is fully comparable to the Walker Mountain (Pulaski) fault. The Saltville fault (Bland fault to the northeast) is 200 miles long from the north end of Sinking Creek valley, Craig County, to a point about 50 miles southwest of the Virginia-Tennessee line. It bounds the Greendale syncline on the southeast. At Greendale, Washington County, the lower part of the Honaker dolomite (Middle Cambrian) is thrust upon Mississippian strata, giving a stratigraphic displacement of about 14,000 feet. Northeastward the fault crosses the Greendale syncline diagonally from west to east, and about 4 miles west of Bland, Bland County, Cambrian dolomite is in contact with Devonian shale. Still farther northeast only Cambrian and Ordovician formations are involved, and

the stratigraphic displacement is probably comparable to that along the Walker Mountain fault in the vicinity of Shortsville (Pl. 1). In fact, the Saltville fault bears the same relation to the rocks of the Greendale syncline as does the Walker Mountain fault to the Walker Mountain syncline, except that the overlap of formations by faulting is in the reverse direction.

MINOR CROSS FAULTS

Approximately 30 cross faults, some of which are slightly oblique, with displacements in the horizontal plane that range from a few feet to as much as 800 feet were identified and mapped (Pl. 1) in the eastern part of the Walker Mountain block. These faults are grouped and well displayed in the vicinity of McCall and Smyth gaps, southwest of Giesler Mill, and on the west side of Walker Mountain southwest of Bald Knob.

The recognition of these cross faults is not difficult where the involved strata are thin and lithologically distinct, but where the faults pass into thick, more or less homogeneous bodies of shale or limestone their identification is uncertain. For this reason, none of the mapped faults are extended on Plate 1 far beyond the limits of observed offset beds. Accurate mapping of the truncated parts of the displaced strata restricts within narrow limits the position of the mapped fault traces. It is inferred that the faults are nearly vertical, because a line connecting points marking the termination of the offset beds is straight or gently curving and apparently unaffected by topographic irregularities. A few observations indicate that the beds on opposite sides of some of the faults vary a few degrees in the amount of dip, whereas many other cross-faulted beds show no variation in dip. It appears likely, therefore, that the displacement is of the strike-slip variety.

In the vicinity of McCall Gap all the cross faults terminate at the north within the thick Martinsburg shale. To the south they extend into the Devonian shale of the overridden block and at certain localities pass into the overthrust sheet of Honaker dolomite. Similar relationships exist at Smyth Gap. Possibly, however, some of the cross faults at the south may extend northwestward through the Martinsburg shale and connect with faults that locally offset the Ottosee and Lowville-Moccasin formations on the north side of Walker Mountain west of Bald Knob. These cross faults appear similar to the cross faults radiating from Opossum Valley thrust west of Birmingham, Alabama, and shown in the Birmingham folio.

CONCLUSIONS

Many of the following interpretations are only hypotheses, for their validity depends upon assumptions which cannot be proved by the writers.

Displacements of large magnitude, such as occurred along the Walker Mountain fault, must have been accompanied by differential movements within the materials both above and below the sole of the thrust which resulted in the development of minor structures. It is thought that cross stresses were developed in the Walker Mountain block as a result of horizontal compression acting unequally on certain parts of the block, thereby being resolved into directed components. Most important is a consideration of why the fault block should be stressed differentially during thrusting. The answer to this question is not known, but a general explanation will

be offered in the hope that it may contribute something to a real solution of the problem. Assuming that the Walker Mountain fault surface was irregular, possibly broadly undulating, then further movement along the sole of the thrust may have encountered areas of locally variable friction. During the early stages of thrusting, it is thought that minor adjustments along the fault surface, possibly the squeezing of the truncated east limb of the Walker Mountain syncline beneath the sole of the fault, plus easy vertical relief would dissipate the horizontal thrusting force so that the rocks in the Walker Mountain block at no great distance west of the fault would be undisturbed. Toward the final period of thrusting we may imagine a greater mass of rock upon the Walker Mountain block. Of still greater importance may be the close parallelism between the dip of the fault surface and the dip of the strata comprising the northwest limb of the Walker Mountain syncline. If these inferred conditions existed, additional pressure probably would have been exerted through the fault surface into the underlying block. Greater shortening would be expected within the block along the northwestward extensions of lines marking zones of maximum friction along the fault surface; less deformation would be expected along lines of minimum friction. The geologic map (Pl. 1) of the southwestern end of the Walker Mountain block shows these probable relationships. The strike of the rocks conforms to a broad recess along a synclinal axis extending northwest of Seven Springs. The rocks in this part of the Walker Mountain block have moved farther northwestward than the same rocks which crop out along the Washington-Smyth county line. Similar relations exist with respect to a line extending northwest of Bald Knob on the southwest.

It is thought that the cross faults in the Walker Mountain area represent shears produced during the warping or the formation of the cross folds by local rotational stresses in the horizontal plane. In the vicinity of McCall Gap, midway between two cross folds, there are 10 cross faults which strike generally north. In each fault the rocks west of the fault trace moved relatively north. Southwest of Bald Knob another group of cross faults occupies similar shearing positions. It is not clear why there are no cross faults along the east slope of Walker Mountain between Smyth Gap and Seven Springs, since this part of the block probably was affected by shearing. It may be that, since the strike of the beds in this area approximates the trend of the principal shear zone, movement took place along the bedding planes.

A few cross faults extend into the thrust mass overlying the Walker Mountain block and locally offset the trace of the Walker Mountain fault. Since the shearing zones within the Walker Mountain block are thought to have been formed during the late stages of overthrusting of the overriding block from the southeast, it is to be expected that some of the cross faults would extend into the overlying thrust mass. Possibly certain adjustments of the rocks, as an aftermath of thrusting, occurred in the area, but it is not thought that they contributed to the origin of the subordinate structures.

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LAMPROPHYRIC DIKE CUTTING MICA SCHIST
MIDDLE OUTCROP (Fig. 4)

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UNUSUAL LAMPROPHYRIC DIKES IN THE MANHATTAN SCHIST OF NEW YORK CITY

BY R. J. COLONY[1] AND HORACE R. BLANK

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ABSTRACT

Unusual lamprophyric dikes, ranging in thickness from a few inches to 10 feet, have been found in outcrops and in shaft and tunnel excavations in the northern part of the Borough of the Bronx, New York City. They cut across the structure of the Manhattan schist and its associated pegmatites.

[1] Deceased.

All the known occurrences can be included in a block about three quarters by one third of a mile in horizontal extent and 600 feet deep. Most of the dikes strike N. 50°–75°W. and dip practically vertically.

The dike rock is massive, bluish gray when fresh, and felsitic. In most of the dikes it is thickly sprinkled with spots of various sizes and colors, which give it a very striking porphyritic appearance. Actually these spots are xenoliths.

Microscopically the rock itself, apart from the xenoliths, is more or less porphyritic. It consists essentially of a few large plates of biotite and many smaller interlocking prisms and needles of apatite and kataphorite, whose interstices are filled with very small equidimensional grains of orthoclase. Accessory minerals are sanidine, aegirite or acmite, arfvedsonite, and sphene. Some of the biotite has altered to chlorite, and a little pyrite and carbonate have been introduced. Other alteration products are a little ilmenite, epidote, and leucoxene. A few granular aggregates may be pseudoleucite.

The xenoliths have been derived from pegmatite, mica schist or gneiss, limestone (now much silicated), amphibolite, serpentine, garnet, and possibly others not easily determined. Most of them have been more or less changed by the magma, which in contrast has had very little effect on the wall rock where the dikes are now exposed.

Modal analyses of the rock from two different dikes are given. The extraordinarily large proportion of apatite (13 to 16 per cent) is a striking feature of both. Corresponding chemical analyses show very low Al_2O_3, high alkalis relative to lime and magnesia, and unusually high K_2O, P_2O_5, and TiO_2. The norms of both lead to the CIPW symbol III,5,1,1.

The rock seems to be unique mineralogically. In most of its characteristics it approaches a minette, but in the typical dikes amphibole predominates over biotite and almost excludes pyroxene. The rock contains no feldspathoids and no plagioclase. It may be named an amphibole-apatite minette.

Related rocks are discussed, and their analyses are given. A possible relation is suggested between the New York City dikes and the nephelite-syenite of Beemerville, New Jersey.

The dikes were intruded after the metamorphism of the Manhattan schist and before the formation of the pre-Cretaceous peneplain of the Atlantic coastal region.

HISTORY AND ACKNOWLEDGMENTS

During the summer of 1923, Mr. R. G. Bergman, a student at Columbia University, discovered an unusual rock cropping out as a dike in the Manhattan schist in the northern part of the Borough of the Bronx, New York City. The late Professor R. J. Colony made a preliminary study of the rock and described it in an unpublished lecture entitled *A basic dike in the Manhattan schist of New York City*, given before the Section of Geology and Mineralogy of the New York Academy of Sciences on March 3, 1924.

In 1927 the Board of Water Supply of the City of New York found a gray, spotted rock in a boring at the site selected for Shaft 3A of City Tunnel No. 2, about two thirds of a mile northwest of the outcrop mentioned. Dr. Charles P. Berkey, consulting geologist to the Board, recognized it as an igneous dike and termed it a feldspathoid porphyry. Microscopic investigation has shown that the rock, although it has feldspathoid affinities, contains no surviving feldspathoids and is not a typical porphyry.

The excavations for City Tunnel No. 2, made in 1929–30, exposed nine or more dikes of this rock and afforded the junior author, then employed as assistant geologist by the Board of Water Supply, a much better opportunity to collect and examine unweathered material and to study the relation of the dikes to the surrounding rocks.[2] His results were discussed with Professor Colony, and the identity of the rock from the outcrops and excavations was then established. Further investigation lagged, however, chiefly because of lack of chemical data. After Professor Colony's death

[2] These studies are more completely described in an unpublished report by the junior author dated January 17, 1934, and entitled *The feldspathoid porphyry, a unique igneous intrusion into the Manhattan schist*, on file in the library of the Board of Water Supply at 346 Broadway, New York City.

in 1936, his notes and material were kindly turned over to the writer by Professor Berkey. Finally, the chemical analyses were made possible in 1938 by a grant from the Penrose Bequest of The Geological Society of America.

Professor Colony must be credited with the first correct understanding of the nature of the rock and with the identification of its peculiar amphibole.

The writer gratefully acknowledges the assistance of the members of the engineering staff of Shaft 3A of City Tunnel No. 2 of the Board of Water Supply in collecting the field data in the shaft and tunnels. He is also indebted to Columbia University and to the Soil Conservation Service for the use of microscopes and office facilities at various times during the course of the investigations. He also wishes to thank Dr. R. B. Ellestad for the difficult and painstaking work required by the chemical analyses, and Drs. C. P. Berkey, C. S. Ross, W. T. Schaller, and E. N. Cameron for their interest and helpful advice.

OCCURRENCE

OUTCROPS

The outcrops in the northern part of the Borough of the Bronx, New York City, lie along the west side of Baychester Avenue about 700 feet south of East 233rd Street (Figs. 2, 3, 4). None of the dikes occurring here is more than 2 feet thick. All follow irregular courses, cutting across the foliation of the schist (Figs. 3, 4, Pl. 1). Their alignment strongly suggests that they occupy intersecting joints which strike N. 5°–25°W. and N. 60°–80°W. Their dips are vertical.

These outcropping dikes all contain numerous more or less altered xenoliths, which however are less conspicuous than in the dikes seen underground. Contact effects are confined to a reddish staining of the feldspars in some of the abutting pegmatites to a depth of perhaps an eighth of an inch.

Another outcrop, located by Professor Colony in 1923 in a small stream gorge on the east side of Baychester Avenue, was not accessible to the writer in 1933.

UNDERGROUND OCCURRENCES

The boring at the site of Shaft 3A (Fig. 2) cut a 5-foot dike of this same rock, dipping 75–80° and rich in small xenoliths. A similar dike was found in a later boring about 100 feet to the northwest.

In the excavation for the valve chamber at Shaft 3A, two parallel dikes, 4 inches and 9 feet wide, were exposed beneath 55 feet of glacial drift. They strike N. 74°W. and dip nearly vertical. At this bedrock surface the dike rock was completely decayed to a stiff yellow clay, which gradually hardened downward and became a bluish-gray rock. It was relatively more decayed than the adjacent schist.

Shaft 3A, excavated to a total depth of 670 feet, passed through a 2–6 foot dike (designated Dike A) striking N. 60°W. and dipping 84°SW. Its vertical section (Fig. 5) suggests that like the dikes in the outcrops it has followed two sets of preexisting joints. The rock composing this dike was hard and undecayed throughout. It contained numerous xenoliths, some as much as 3 to 4 feet across. Those of the gray feldspar (pegmatite) type were particularly prominent. The wall rock showed no contact metamorphism beyond the slight reddish staining of the feldspars noted

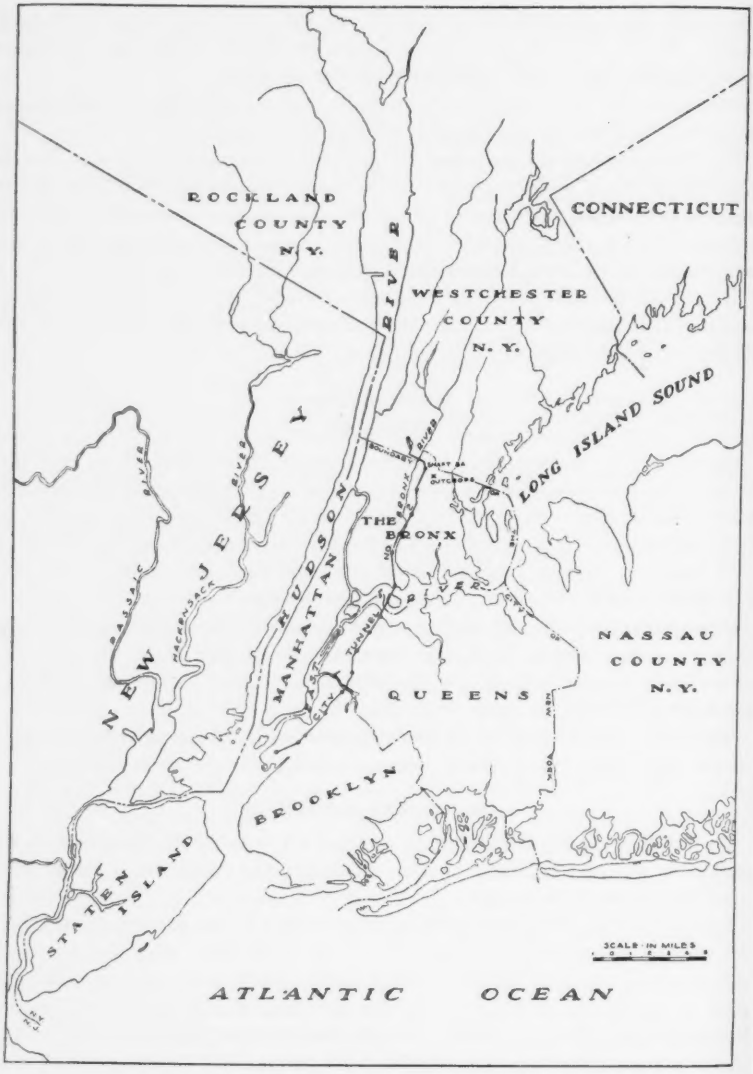

FIGURE 1.—*Index map of New York City and vicinity*

in the outcrops. In some places a very thin hard green band separated the dike from the wall rock. The dike rock nowhere penetrated between the foliation planes of the schist.

The tunnel, driven nearly horizontally in two directions from the foot of the shaft,

cut nine distinct and more or less parallel dikes in about a third of a mile. The more important are shown in Figure 2. The largest averaged 10 feet thick, and the smallest less than one inch. Their strikes ranged from N. 47°W. to N. 80°W., and their dips were within 15° of the vertical. Some of the dikes were slightly curved; others forked within the limits of the tunnel (Pl. 2, fig. 1).

FIGURE 2.—*Map showing relation of outcrops, shaft, and tunnel*

All the dikes cut directly across the structure of the schist and pegmatite, and none penetrated the schist along its foliation. Small displacements of the schist-pegmatite contacts suggested that the lamprophyric dikes had filled minor faults. Contact effects were limited to the slight red staining of the feldspars previously mentioned.

The texture of the dikes varied with their thickness. The proportion of xenoliths, however, showed little relation to the size of the dike, and only the smallest dikes were entirely free of them. Some of the dikes were extensively jointed.

Dike #2 in the tunnel, approximately a foot thick, differed from all the others in having chilled borders (faintly seen in Pl. 2, fig. 2) showing flow structure. The contacts effects were a little more pronounced at this dike.

CORRELATIONS

From their many common features it seems safe to conclude that all the dikes described above are offshoots from the same igneous mass. Figure 2 shows the rela-

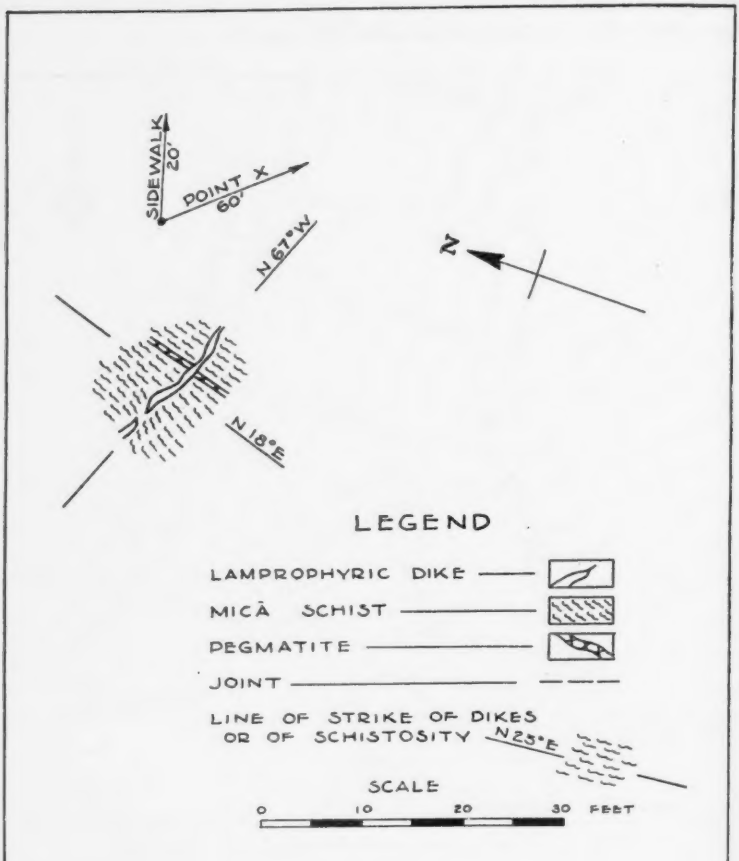

FIGURE 3.—*Map of outcrops of lamprophyric dikes (northern portion)*

tion between the dikes along Baychester Avenue and those in the tunnel. Apparently the outcrops lie almost in line with the largest dike in the tunnel, and well within a zone delimited by the outermost dikes, which are practically parallel.

DESCRIPTION OF THE ROCK

GENERAL APPEARANCE

The unweathered rock from the excavations is massive, bluish-gray, and fine-grained. On the slightly weathered surface of the glaciated outcrops it is darker and

duller. Its texture, although varying somewhat, is everywhere felsitic. The rock is fairly hard, although it can be marked with a knife, and is in general extremely difficult to break except along natural joints.

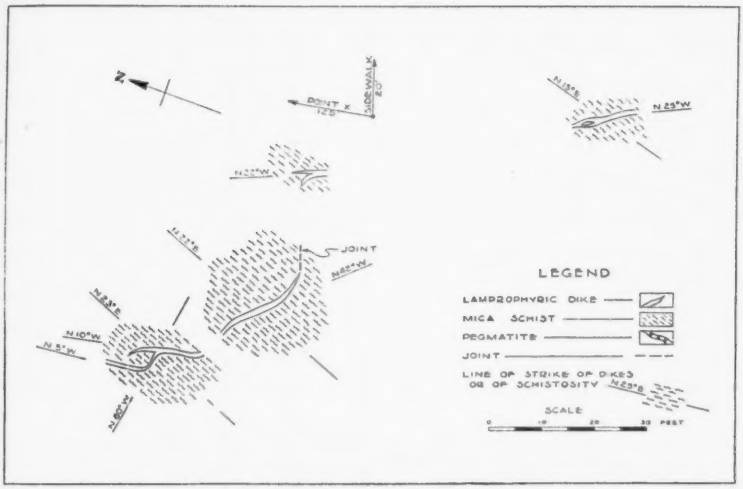

FIGURE 4.—*Map of outcrops of lamprophyric dikes (southern portion)*

In most of the dikes the felsitic groundmass is thickly sprinkled with small and large spots of several different colors, which give the rock a striking porphyritic appearance, but which are in reality more or less digested xenoliths. The abundant, very small, true phenocrysts of biotite are revealed by the somewhat glittering appearance which they impart to a broken surface.

TEXTURE, MINERAL COMPOSITION, AND STRUCTURE

Microscopic examination shows the rock to be entirely crystalline. The primary essential minerals are biotite, apatite, the amphibole kataphorite, and orthoclase feldspar, stated roughly in order of decreasing grain size and in inverse order of abundance.

The structure is in general massive. It consists essentially of a few large plates (biotite and some feldspar) and many smaller interlocking prisms and needles (apatite and kataphorite), whose interstices are filled with small equidimensional grains (feldspar). Grain sizes vary from particles only a few thousandths of a millimeter in dimensions to prisms about 0.4 millimeter long and plates about 0.6 millimeter across. Nearly all sizes within this range are represented, but the larger plates and prisms are much less abundant, so that in thin section the rock has a somewhat porphyritic appearance, which is much more marked in some specimens than in others. In this respect, and in the relative proportions of the essential minerals, the thin sections examined may be divided into four groups. In seven sections representing various dikes the kataphorite exhibits a wide range of sizes, from prisms as large as those of the apatite down to very tiny grains, and thus the rock is not prom-

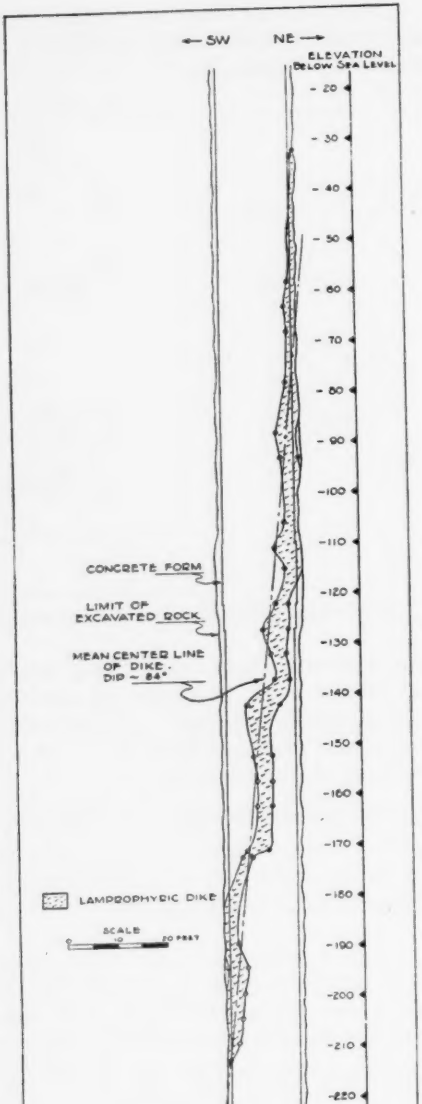

FIGURE 5.—*Vertical section of dike in Shaft 3A*

inently porphyritic. In contrast, five sections, equally widely distributed, show quite a hiatus in grain size between the smaller phenocrysts and the groundmass, which consists of short, stumpy prisms of kataphorite interspersed with very fine-grained feldspar. One section, from Dike 2 in the tunnel, is unique in containing acmite instead of kataphorite, in the abundance of chlorite, and in showing indications of flow structure. Two sections, from the stream gorge outcrop, differ from all the others in having no biotite.

PRIMARY ESSENTIAL MINERALS

The biotite occurs as large and small plates and short prisms, in places of perfect hexagonal outline. They are reddish-brown on the base but seen edgewise they are pleochroic from pale yellow to bright yellowish-brown, suggesting a low iron content. In places they are invaded by feldspar and kataphorite. A few crystals show zonal growth; some show partial alteration to chlorite (near center of figure 2 of Plate 3); others show also partial alteration to a dull black, opaque mineral (possibly ilmenite).

Apatite forms long prisms, which average somewhat smaller than the biotite crystals. They are more or less hexagonal in cross section, and some have blunt pyramidal terminations (Pl. 3, fig. 3). They are clear, colorless, unaltered, and contain no inclusions. The apatite is unusually abundant, averaging about 13 per cent of the rock by volume.

Kataphorite, an amphibole between hornblende and arfvedsonite and rich in sodium and iron, occurs as a network of poorly formed prisms and needles of various sizes, constituting about 40 per cent of the rock (Pl. 3, fig. 2). Most of its prismatic sections show rather large extinction angles up to a maximum of 34 degrees, with moderate pleochroism in shades of greenish-yellow, yellowish-green, and bluish-green, and no marked absorption. The elongation varies from positive to negative in different grains, and even within the same grain. Sections showing small extinction angles, however, and those rare sections which show parallel extinction, are strongly pleochroic from greenish-yellow parallel to the prism to deep violet or violet-brown perpendicular to it, with strong absorption in this latter direction. The occasional hexagonal and diamond-shaped cross sections, which show good amphibole cleavage, show the violet color and strong absorption in the direction of the *b* axis.

The interstices between the grains of the preceding minerals are filled almost entirely with small, irregular grains of orthoclase, only a few of which show any crystal edges. Much of this feldspar is somewhat cloudy or "dusty".

PRIMARY ACCESSORY MINERALS

Perfectly formed phenocrysts of sanidine occur in some of the thin sections (Pl. 3, fig. 4). They have a square, rectangular, rhombic, or irregularly octagonal outline and are perfectly clear and unaltered. Their abundance seems to vary greatly, even in adjacent parts of the same dike.

A few large, irregular grains of an undetermined feldspar, much corroded and altered, were also observed (Pl. 3, figs. 1, 4). These may be either phenocrysts of a still earlier feldspar generation or xenocrysts from a pegmatite.

Aegirite, as needles grown into and on the feldspar and the kataphorite, is abundant in and near certain xenoliths, but elsewhere it appears to be only a very minor constituent. It is not easy to distinguish from the kataphorite, but almost all the aegirite needles are smaller and thinner. They show nearly parallel extinction, only slight pleochroism from deep green to yellowish-green, and do not possess the characteristic absorption of the kataphorite.

In the thin section from Dike 2 no kataphorite is present, but clusters of small, greenish-yellow prisms, in places radially arranged, are fairly abundant. These have all the optical properties of aegirite except the usual deep green color and are probably acmite (Pl. 3, fig. 3). Extremely small, thin, green aegirite needles are also abundant in this thin section.

The thin sections from the stream gorge outcrop contain no biotite and less apatite than the average, but the kataphorite is very abundant and reaches larger than average size. Many of its larger prisms and blades show longitudinal streaks of arfvedsonite, pleochroic from blue to greenish-yellow.

Numerous small, irregular grains of sphene, scattered throughout the rock, occur in many of the thin sections.

Neither quartz nor olivine was found anywhere in the rock.

SECONDARY MINERALS AND ALTERATION PRODUCTS

Small quantities of chlorite occur as an alteration product of the biotite in most of the dikes, but nearly all the biotite shows some of this alteration in the thin section from Dike 2. Chlorite also occurs here in groups of radiating narrow blades, whose structure suggests a possible derivation from the acmite present in this dike.

Fine granular, yellowish epidote occurs in this same thin section as narrow zones and patches around many of the corroded biotite grains. It was not observed in thin sections from the other dikes.

Ilmenite is quite abundant in some of the thin sections, most of it being directly associated with biotite. In the thin section from Dike 2 the ilmenite and much of the sphene are coated with leucoxene.

Small, symmetrically rounded areas of low relief appear in several of the thin sections as colorless "clearings" in the tangled growth of kataphorite and feldspar (Pl. 3, fig. 1, lower left). Their boundaries are irregular and corroded, and many are invaded by the kataphorite. Between crossed nicols they are seen to be aggregates of irregular, very small, weakly birefringent grains. Their more or less circular form suggests former phenocrysts of leucite, which have now been altered to pseudoleucite, an aggregate of orthoclase and nephelite. These aggregates are not common and are the only suggestion of the possible former presence of a feldspathoid mineral.

A few small, irregular patches of carbonate occur. Here and there pyrite, in larger patches and isolated cubes, invades and replaces all the other minerals.

All the thin sections made from outcrops show iron staining in small patches and as a general cloudiness. This alteration is entirely absent in the unweathered rock from the borings and excavations.

XENOLITHS

The abundant xenoliths form by far the most conspicuous feature of the dikes. They range from 3 feet or more across down to grains comparable in size to the

FIGURE 1. DIKE 5 and BRANCH CUTTING INJECTED MICA SCHIST
Note xenoliths in dikes and pegmatite lenses in schist. Large xenolith to right of center is about 8 inches across. Roof of tunnel. N.Y.C.B.W.S. photograph.

FIGURE 2. DIKE 2, SHOWING CHILLED EDGES.
Each arm of rule is foot. N.Y.C.B.W.S. photograph.

TUNNEL EXPOSURES

FIGURE 1. DIKE A
Lower nicol only. X 46. Altered feldspar, upper right; Possible pseudoleucite, lower left.

FIGURE 2. DIKE A
Lower nicol only. X 46. Biotite altering to chlorite, right center; Kataphorite, small, moderately dark grains, scattered.

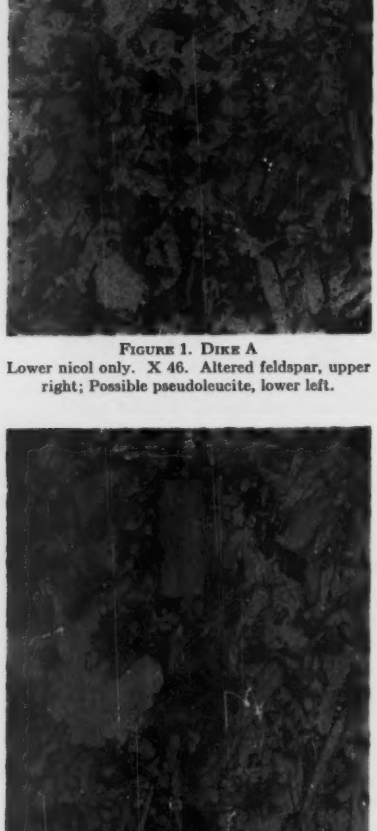

FIGURE 3. DIKE 2
Lower nicol only. X 46. Biotite, upper center; Acmite, in clusters near center; Apatite, long prisms.

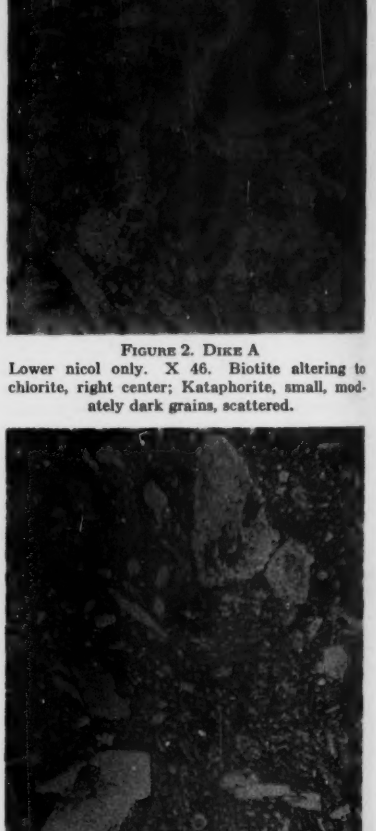

FIGURE 4. SPECIMEN FROM TUNNEL
Lower nicol only. X 46. Sanidine phenocrysts, lower left; Altered earlier feldspar, upper right.

PHOTOMICROGRAPHS

larger phenocrysts. Except in some of the narrower dikes, it is difficult to obtain a hand specimen which does not contain at least a few. They are of many different types, but most of them have been more or less modified by the surrounding magma, so that it is impossible in many cases to identify their origin with any certainty.

Xenoliths of pegmatite are gray and average 6 inches or more in diameter. They are very coarse grained, rounded, and generally structureless. Their chief constituent is bluish-gray feldspar, apparently sodic sanidine, associated with considerable gray quartz and a little plagioclase and biotite. They do not represent the pegmatites which compose part of the wall rock of the lamprophyric dikes. The latter pegmatites are white or pinkish and of slightly different mineral composition and do not carry the characteristic bluish-gray feldspar of the xenoliths.

Some xenoliths consist of biotite schist; others are more gneissic and contain feldspathic bands which show greenish and pink stains like those already noted as contact effects in the wall rock.

A large number of xenoliths appear to represent fragments of limestone which have been silicated and perhaps partly digested by the lamprophyric magma. These are white, pink, or red, or mottled with these colors. They vary in size, shape, texture, and structure, but most are well rounded, massive, and less than 6 inches long. They consist chiefly of feldspar and calcite, with a little quartz. The calcite, originally in large grains, has been invaded and partly replaced by the feldspar, and minute green needles of aegirite have later grown in both. The feldspar, which includes both orthoclase and plagioclase, is all more or less kaolinized, and in the red xenoliths much of this kaolinized material is pink to reddish. A little wollastonite, graphite, and octahedrite are present in some specimens.

At this locality the Manhattan schist should be underlain at depth by the Inwood limestone, and this in turn by the Fordham gneiss. Both schist and gneiss are known to contain interbedded layers of impure limestone near by.

Amphibolite xenoliths are generally not over 2 inches across, irregular in shape, medium- to fine-grained, and for the most part very dark and massive. They are composed of about 75 per cent hornblende, the remainder being biotite, quartz, and feldspar. They may represent either silicated very impure limestones or hornblende schists, and it seems probable that the dike magma has not greatly changed their original composition.

Serpentine xenoliths are mostly less than an inch long, oblong to very imperfectly rounded, fine- to medium-grained, and massive. They are dark green and rather friable. There is commonly a dark stain in the surrounding dike rock, and a thin line of very fine-grained biotite at the contact. Nearly two thirds of their material is brown serpentine (both antigorite and chrysotile), and the remainder chiefly pale-green diopside, with a little quartz and calcite.

The origin of the serpentine is obscure. The association suggests derivation from the diopside, but the latter is fresh and unaltered. An intrusion of serpentine occurs at New Rochelle, New York, a few miles away, and amphibole occurs with diopside in many of the bands of metamorphosed impure limestone in the Manhattan and Fordham formations.

A few large, well-rounded, grayish-green xenoliths are not conspicuous in the dark dike rock. These consist almost entirely of feldspar, chiefly andesine, much of which

encloses tiny dark-green prisms of aegirite and kataphorite. They show broad, indefinite "reaction rims" of slightly lighter color. These xenoliths may represent completely silicated limestones, although they now contain only traces of carbonate.

One greenish xenolith, similar in mineral composition to the red and white limestone ones, contains a lens rich in brownish-yellow barite.

Small, irregularly rounded spots (seldom over one quarter of an inch across) occur within many of the white limestone xenoliths, also in a few of the red ones, and rarely as independent inclusions in the lamprophyric dikes. In some specimens they straddle the contact between xenolith and igneous rock, as if formed by the replacement of both. They are dull, greenish black, fine-grained, and massive. Some contain visible grains of garnet.

Microscopic study shows that these spots are former garnets (apparently individual crystals originally), now more or less completely altered to chlorite at their margins and along all their fractures. The unaltered garnet is slightly purplish and is perfectly isotropic. They appear to have been connected with some reaction between the dike rock and the limestone xenoliths, but their occasional occurrence alone is not easily explained.

CLASSIFICATION AND PETROLOGIC RELATIONS

MINERALOGICAL CHARACTERISTICS — THE MODE

An attempt was made to determine the mode on the thin section made from a chemically analyzed specimen from Dike A in the shaft, which is representative of the typical dike rock free from xenoliths. The only equipment available for this purpose was a 10× ocular containing a scale of 100 divisions. Using a 10× objective in connection with this, the thin section was moved about at random on the stage of the microscope, and at each position the nature and width of the mineral grains which happened to lie along the center of the scale were noted.

Similar mineral counts were made on the thin section from an analyzed specimen from Dike 2. The results of these counts are given in Table 1.

The crystals of sanidine mentioned under Description of the Rock were not present in these thin sections, and none of the very few aggregates of pseudoleucite fell in the cross sections used. A little aegirite fell on the scale in the thin section from Dike 2, but the needles were so small that no attempt was made to measure them.

As may be noted, the rock is intermediate in composition, and there is no great excess either of feldspar or of ferromagnesian minerals, although the latter predominate slightly. Despite this, none of the feldspar is plagioclase. There is no quartz, but neither are there any feldspathoids, although both mineralogical and chemical analyses suggest that the rock has feldspathoid affinities. Apatite is extraordinarily abundant.

According to the classification of Johannsen (1931, p. 140–158), the rock, having dark minerals between 50 and 95 per cent by volume and neither quartz nor feldspathoids, would fall into his Class 3, Order 1, and exactly on the dividing line between Family 9 and Family 13. Assuming that the rock has feldspathoid affinities it should probably be placed in Family 13. This makes its number 3–1–13 H (hypabyssal), but it is not far removed from 3–1–9, 2–1–13, and 2–1–9.

CHEMICAL CHARACTERISTICS — THE NORM

Chemical analyses of two specimens of the dike rock were made by Dr. R. B. Ellestad at the Laboratory for Rock Analyses of the University of Minnesota. His results are given in Columns I and II of Table 2. A letter accompanying them mentions that:

"The high content of zirconium is unusual, especially since no zircon was observed.... The strontium and barium content is also unusually high.... Some of the sulphate is acid soluble, while some is insoluble, although the proportions of each were not determined."

TABLE 1.—*Modes of lamprophyric dikes, New York City*
(Per cent by volume)

	Thin section from Dike A	Thin section from Dike 2
Number of cross sections counted	20	10
Orthoclase	35	30
Apatite	13	14
Kataphorite	40	0
Biotite	8	18
Acmite	0	8
Sphene	3	6
Ilmenite	1	1
Chlorite	0.1	18
Epidote	0	5
Total	100.1	100
Light minerals	48	44
Dark minerals	52	56

Great care was exercised in selecting the specimens for analysis to obtain nothing but the dike rock and to avoid any visible inclusions, although inclusions were so numerous in Dike A that this effort may not have been entirely successful.

The norm of the rock in the CIPW system, calculated according to the rules and tables given by Johannsen (1931, p. 83–99, and appendices), is shown in Table 3.

The calculation of the norms emphasizes what may be observed in the analyses—that it is not silica but alumina which is strikingly deficient in this rock. The alkalis are very high and must enter into the alkali-iron amphibole and the mica. The great excess of potash over soda is striking, and the calcium is little more than sufficient to take care of the phosphorus and the carbonate. These facts are reflected in the absence of plagioclase and of any pyroxenes or amphiboles which are prominently calcium-bearing. The magnesium is probably largely in the phlogopitic biotite.

The content of titania is also extraordinarily high in both analyses, and in view of the comparative shortage of calcium and ferrous iron it does not seem possible to account for all the titanium as sphene or ilmenite. Some of the grains considered as sphene in the modal analysis may be rutile, but no prismatic crystals were observed. It seems probable that titanium enters into both the mica and the amphibole. The zirconia likewise is not present as zircon and probably is in the kataphorite and ac-

TABLE 2.—*Chemical analyses*

	I	II	1	2	3	4	5	6	7	8	9	10	11	12	13	14
SiO_2	49.73	46.75	50.41	46.04	32.32	41.57	55.56	51.22	51.95	40.71	51.16	54.08	50.78	47.19	50.00	31.80
Al_2O_3	9.93	8.64	12.27	12.23	8.16	9.75	10.70	17.56	14.95	19.46	14.25	9.49	9.05	23.01	20.03	18.78
Fe_2O_3	2.89	3.80	5.71	3.86	9.46	4.06	2.00	3.51	4.09	7.46	3.76	3.19	2.96	3.11	.98	15.20
FeO	3.01	2.22	3.06	4.60	4.10	4.47	5.19	4.34	5.70	6.83	4.39	1.03	3.64	2.23	3.98	n.d.
MgO	5.85	7.11	8.69	10.38	5.97	8.65	4.56	3.22	3.54	6.21	6.09	6.74	14.29	1.07	.69	3.32
CaO	5.71	7.09	7.08	8.97	12.60	11.10	8.42	4.52	6.10	11.83	6.37	3.55	5.29	2.93	3.41	14.60
Na_2O	1.93	1.05	.97	2.42	.69	1.57	1.45	5.72	5.43	1.80	2.40	1.39	1.05	7.97	8.28	1.10
K_2O	10.03	9.37	7.53	5.77	5.97	6.10	7.48	4.37	4.45	3.26	4.95	11.76	7.39	8.23	8.44	5.07
H_2O+	.72	1.48	1.80	2.87	4.09	2.30	1.45	1.93	1.10	1.53	2.75	2.71	2.53	.55	1.50	8.10
H_2O-	.10	.29	.46		1.03	1.54	1.42					.79		.04	.10	
CO_2	.41	.87			6.30	1.24		.60		.74	1.30		.46	.38	.22	
TiO_2	3.04	3.99	1.47	.64	4.55	2.36	.81	1.70	1.95		1.36	2.08	1.28	2.16	.99	
ZrO_2	.28	.40	none			.02								.13		
P_2O_5	3.54	4.98	.46	1.14	3.78	4.05		1.08	1.15		.76	1.35	1.31	.39	.21	.95
SO_3	.58	.25										.29		.45		
Cl	.02	.02	trace	.11		.04						.04		.01	trace	
F	.62	.64				.23						.49		p.n.d.		
MnO	.11	.10	.15	trace	.13	.25	.61	.20	.30	.18	.21	.05	.09	.16	.50	
SrO	.37	.55	.06	.25	.24	.11						.20				
BaO	1.13	.34	.23	.48	.36	.44						.67		.09	none	
FeS$_2$	.28	.19				.06									.54	
Others			.07		.26	.10					.30	.07		.06		
Total	100.28	100.13	100.42	99.76	100.01	100.01	99.65	99.97	100.71	100.01	100.05	99.97	100.12	100.16	99.87	98.92
Less O = F, Cl, S	.26	.27		.03	.13	.11						.22				
Total	100.02	99.86		99.73	99.88	99.90						99.75				
Specific gravity	2.861	2.857	2.88				2.67					2.686			2.701	

I. Amphibole-apatite minette, Borough of the Bronx, New York City. Dike A. R. B. Ellestad, analyst. Specific gravity determined on chips by H. R. Blank.

II. Amphibole-apatite minette, Borough of the Bronx, New York City. Dike 2. R. B. Ellestad, analyst. Specific gravity determined on chips by H. R. Blank.

1. Prowersite (syenitic lamprophyre), Two Buttes, Colorado. W. F. Hillebrand, analyst. (Cross, 1906, p. 168).
2. Cascadite, Arrow Peak, Highwood Mountains, Montana. H. W. Foote, analyst. (Pirsson, 1905, p. 145).
3. Minette, Apishapa Quadrangle, Colorado. G. Steiger, analyst. (Cross, 1915, p. 19).
4. Apatitic minette, near Northport, Columbia River, Washington. W. F. Hillebrand, analyst. (Ransome, 1908, p. 338).
5. Augite minette, Studencho, near Eule, Bohemia. J. Friedrichsen, analyst. (Rosicky, 1901, p. 29).
6. Soda minette, Brathagen, Norway. V. Schmelk, analyst. (Brögger, 1897, p. 130).
7. Soda minette, Håo, Norway. V. Schmelk, analyst. (Brögger, 1897, p. 139).
8. Minette, Franklin Furnace, New Jersey. L. G. Eakins, analyst. (Iddings, 1898, p. 238).
9. Mean of 32 minettes. (Johannsen, 1937, p. 36).
10. Orendite, Fifteen-Mile Spring, Leucite Hills, Wyoming. W. F. Hillebrand, analyst. (Cross, 1897, p. 130).
11. Jumillite, Jumilla, Murcia, Spain. M. Dittrich, analyst. (Osann, 1906, p. 290).
12. Nephelite-syenite, Beemerville, New Jersey. M. Aurousseau, analyst. (Aurousseau and Washington, 1922, p. 576).
13. Leucite tinguaite, Beemerville, N. J. J. E. Wolff, analyst. (Wolff, 1902, p. 276).
14. Ouachitite, Beemerville, N. J. J. F. Kemp, analyst. (Kemp, 1889, p. 133).

mite. The high content of barium and strontium suggests the presence of barium-feldspars.

The presence of considerable SO_3, some of which is acid-soluble as reported by Doctor Ellestad, suggests the presence of feldspathoid minerals of the sodalite group. These were not observed, but no microchemical tests were made.

One of the most striking features of the rock is its extraordinarily high content of apatite. It may be noted, however, that there is not sufficient P_2O_5 in either chemical

analysis to account for all the apatite observed in the modal analysis of the same specimen. A partial chemical analysis made for Professor Colony on a specimen from the outcrops in 1925 gave a P_2O_5 content in reasonable agreement with the figures

TABLE 3.—*Norms of lamprophyric dikes, New York City*

	Dike A (*per cent*)	Dike 2 (*per cent*)
Apatite (ap)	8.40	11.76
Thenardite (th)	0.99	0.43
Ilmenite (il)	5.78	4.71
Rutile (ru)	0	1.52
Fluorite (fr)	0.62	0.43
Calcite (cc)	0.90	1.20
Sodium carbonate (nc)	0	0.85
Zircon (Z)	0.37	0.55
Orthoclase (or)	53.93	47.26
Potassium metasilicate (ks)	1.54	2.31
Acmite (ac)	8.32	2.77
Sodium metasilicate (ns)	0.73	0
Hematite (hm)	0	2.88
Diopside (di) calcium metasilicate	1.51	0
Diopside (di) magnesium metasilicate	1.20	
Diopside (di) iron metasilicate	0.13	
Hypersthene (hy) magnesium metasilicate	11.90	17.80
Hypersthene (hy) iron metasilicate	0.53	0
Olivine (ol) magnesium orthosilicate	1.05	0
Olivine (ol) iron orthosilicate	0.10	
Quartz (Q)	0	2.94
Pyrite (pr)	0.28	0.19
Total	98.28	97.60
Salic minerals	55.29	52.03
Femic minerals	42.99	45.57
CIPW symbol	III,5,1,1	III,5,1,1

shown in Table 2, and in this case also the P_2O_5 was insufficient for the apatite which he observed. Even allowing for the variation in the rock and for the possibly considerable errors in the modal analyses, the fact that this discrepancy is always in the same direction may be significant. The recent studies by McConnell (1937; 1938) and others, which show that other groups may substitute for the PO_4 in apatite, are suggestive in this connection.

RELATED ROCKS

Probably the closest relative of the New York City rock is the "syenitic lamprophyre" from two laccoliths at Two Buttes, Prowers County, Colorado (Table 2, no. 1), described as prowersose by Cross (1906, p. 165–172) and renamed prowersite by Rosenbusch (1922, p. 505). A little less similar is the cascadite of the Highwood

Mountains of Montana (Table 2, no. 2) described by Pirsson (1905, p. 142–149). Cross (1915, p. 17–20) also described a somewhat altered minette from the Apishapa quadrangle of southeastern Colorado (Table 2, no. 3), for which he calculates the same CIPW symbol as that of the New York City rock. A minette very similar to the Apishapa example and notable for its extraordinarily high content of apatite was described by Ransome (1908) from near Northport in northeastern Washington (Table 2, no. 4).

The chemical analysis of the New York City dike rock is strikingly similar to that of the orendite of the Leucite Hills of Wyoming (Table 2, no. 10) described by Cross (1897, p. 115–144, especially p. 123–126), which has the same CIPW symbol. The orendite, however, is an extrusive and contains leucite and sanidine in about equal proportions; these predominate over phlogopite and diopside, the other essential constituents. The jumillite (Table 2, no. 11) from Jumilla, Spain (Osann, 1906, p. 283–301), is similar to orendite in both composition and manner of occurrence.

Washington, in his tables of chemical analyses of igneous rocks (1917, p. 595), gives the analysis of an "augite-minette" from Studencho, near Eule, Bohemia (Table 2, no. 5) (Rosicky, 1901, p. 29), which has the CIPW symbol III,5,1,2.

Other rocks of interest in comparison with the New York City dikes are two "soda-minettes" from Norway (Table 2, nos. 6,7) (Brögger, 1897, p. 126–144), also a kvellite from the Oslo region, Norway, and a jacupirangite from Tjose, Norway (Brögger, 1933, p. 78–80). The kvellite carries the amphibole barkevikite as its principal mafic constituent, but the other rocks contain the more usual pyroxene.

CLASSIFICATION AND NAME

The dike rock is difficult to classify properly and to name, as it differs in mineralogy from all dike rocks of similar chemical composition which have come to the writer's attention. The original field designation "feldspathoid porphyry" is hardly suitable, for feldspathoid minerals are not and never were present in more than very minor amounts at best. In contrast, apatite is so abundant as to be a major constituent rather than an accessory. This is true in all the thin sections examined.

There seems little doubt that the rock should be classed with the lamprophyres, although in the tables of lamprophyres given by Beger (1923) no rocks show quite the same mineral assemblage. In general the New York City rock approaches a minette or a vogesite. It is somewhat porphyritic, at least in certain dikes, with biotite forming most of the phenocrysts, but the feldspar, rather than the dark minerals, occasionally occurs in two generations. Amphibole preponderates over biotite, as in the vogesites; on the other hand, plagioclase is entirely absent, and the alkalis are even higher relative to lime and magnesia than in most minettes. The potash considerably exceeds the entire range given for the 32 minettes averaged by Johannsen (1937, p. 36).

Rocks so rich in alkalis and yet containing no feldspathoids are few. In the most closely related rocks pyroxene is the ferromagnesian mineral accompanying the biotite or phlogopite, and only in the orendite and jumillite does a kataphoritic amphibole begin to appear in very minor quantities. In the typical New York City rock, such an amphibole dominates and almost excludes pyroxene, although, as shown by Dike 2, the balance between the two must be rather delicate. The much more

mafic kvellite is the only other alkali-feldspar lamprophyre in which amphibole is the dominant ferromagnesian mineral. The uniformly high apatite content of the New York City dikes is also very unusual, although not absolutely unique. These features make it difficult to include the rock under even such a general name as minette.

An entirely new name might be justified but is admittedly undesirable. It seems better to resort to the use of mineral qualifiers and to call the New York City dike rock an amphibole-apatite minette.

POSSIBLE RELATION TO ALKALIC ROCKS AT BEEMERVILLE, NEW JERSEY

A relationship is naturally suggested between the New York City dikes and the only other highly alkalic rocks in the surrounding region—the nephelite-syenite of Beemerville, New Jersey, and its associates (Emerson, 1882; Kemp, 1892; Aurousseau and Washington, 1922, p. 571–586). Analyses of the nephelite-syenite, as well as of a small dike of leucite-tinguaite from the same locality described by Wolff (1902, p. 273–277), are given in Table 2 (nos. 12 and 13). The Beemerville rocks are rich in feldspathoids and contain relatively much more alumina, lime, and soda than the New York City dikes. The relatively high zirconia content of the nephelite-syenite is pointed out by Aurousseau and Washington, and this is even more prominent in the New York City rock.

A minette from Franklin Furnace, New Jersey, about 9 miles southeast of Beemerville, is also included in Table 2 (no. 8).

To the south and southeast of the main intrusion at Beemerville a number of bosses and dikes of fine-grained rock occur, which Kemp (1889, p. 130) described as porphyrite or ouachitite (Table 2, no. 14) and believed to be related to the nephelite-syenite. Some of these contain leucite (Kemp, 1894), p. 339), and others analcite, but their feature of greatest interest to the present discussion is that the abundance of apatite impressed Kemp as "one of the most remarkable features of the rock", although the highest P_2O_5 content which he gives—0.95 per cent—is small by comparison with the New York City dikes.

It is interesting to find that a line drawn on the map to connect Shaft 3A of the New York City tunnel and the Beemerville outcrop—about 51 miles apart—has a direction N. 62°W., which is quite close to the average strike of the New York City dikes. The latter therefore conform to the arrangement of the fine-grained alkalic dikes occurring to the southeast of Beemerville and in the vicinity of Franklin Furnace, which are radially disposed toward the main Beemerville intrusion (Wolff, 1908, p. 12), although the nephelite-syenite itself is elongated in quite a different direction (about N. 35°E). A relation between the amphibole-apatite minette of New York City and the nephelite-syenite of Beemerville is somewhat more definitely suggested by these facts.

AGE

The geologic age of the intrusion of the amphibole-apatite minette is of some interest, but unfortunately the evidence indicates only that the dikes are younger than the schist with its *lit-par-lit* injections and also younger than the broader, coarser pegmatites. The pegmatites appear to be at least a little younger than the schist.

None of the lamprophyric dikes show any indication of metamorphism or any tendency toward an imposed schistosity; neither do they show any distortion in harmony with the tightly compressed folds which affect the schist and possibly the pegmatites as well, but they give every evidence of having intruded the schist after after all these structures had been formed. Furthermore, they contain xenoliths of schist and pegmatite, and others apparently derived from the underlying limestone. Nevertheless, at the most northerly outcrop, the foliation of the schist changes direction and passes between the two parts of the intruding dike (*see* Figure 3), in a manner which suggests that the lamprophyre may have been intruded very soon after the metamorphism and folding of the schist and while it was still somewhat plastic.

The geologic age of the rock which is now the Manhattan schist and the time of its metamorphism into schist are much-discussed questions (Merrill, *et al.*, 1902; Berkey and Rice, 1919; Balk, 1936). The rock may be either pre-Cambrian or Paleozoic, and the metamorphism may have occurred in pre-Cambrian time or during the Appalachian Revolution. In any event the lamprophyric dikes must have entered the schist after its metamorphism and before the great pre-Cretaceous peneplanation which truncates all these rocks. If the dikes are related to the Beemerville nephelite-syenite, which is believed to intrude the Shawangunk conglomerate, they are younger than lower Silurian.

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CORRELATION OF THE CENOZOIC FORMATIONS OF THE ATLANTIC AND GULF COASTAL PLAIN AND THE CARIBBEAN REGION

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INTRODUCTION

This is Number 12 of a series of correlation charts prepared under the auspices of the Committee on Stratigraphy of the National Research Council (Dunbar, 1942, p. 429–434). It consists of two units, the first panel, by Cooke and Gardner, showing correlations for the Atlantic and Gulf Coastal Plain and the second, by Woodring, dealing with the Caribbean region. This separation was made because there are still too few key horizons to permit detailed interprovincial correlations.

Both panels of the chart are laid out to correspond with a generalized standard of reference agreed upon by the subcommittee for both eastern and western North America, but the authors of the present chart wish to emphasize that there is little basis for an exact correlation between the Cenozoic formations of the Atlantic and Gulf Coastal Plain and the Caribbean region on the one hand and the European standard section on the other.

The panel covering the West Indies and Central America includes only the islands and countries in which a fairly complete Cenozoic section is represented or is known. The stratigraphic columns are composite and are not representative for any area, particularly in an island as large as Cuba, where the upper Eocene Principe formation, the upper Oligocene Cojimar formation, and the lower Miocene Güines limestone are transgressive.

European stage names are omitted in the West Indian and Central American part of the chart, as their use implies a refinement of intercontinental correlations not now justified. The compiler prefers to assign the supposed equivalents of the

European Aquitanian and the supposed equivalents of the European Pontian (or upper Sahelian) to the Miocene. Alternative assignments to the Oligocene and Pliocene, respectively, are indicated by broken lines at the left of the chart.

Important key faunal zones are shown in the second panel of the chart, the upper and lower limits of each zone being marked by distinctive dotted lines in each column where it has been recognized.

Though the second panel of the chart is published with the approval of the Director of the U. S. Geological Survey, the formation names and age assignments have no official status, as they represent regions not considered officially by that organization.

The following annotations are numbered to agree with numerals on the chart. They deal principally with interpretations different from those of the compilers, and with unpublished data.

ANNOTATIONS

1. This chart, published by permission of the Director of the Geological Survey, Department of the Interior, has been compiled from authentic published sources and from the original investigations of the authors. C. Wythe Cooke prepared the part dealing with the Pleistocene epoch in the Coastal Plain and is mainly responsible for the columns of States east of the Mississippi River. Julia Gardner is mainly responsible for the columns of pre-Pleistocene formations west of the Mississippi River. Wendell P. Woodring prepared the second panel, dealing with the Caribbean region.

2. Flint (General, 1940, p. 773–777) recognizes only two Pleistocene marine shore lines, those bounded by the Surrey Scarp (the shore line at 25 feet above present sea level) and the Suffolk Scarp (the shore line at 100 feet above sea level). He thinks that the evidence for the other five is not conclusive. *See* Cooke (General, 1941).

3. Under previous usage by the U. S. Geological Survey the Iowan drift was regarded as representing a distinct stage of glaciation separated from the Wisconsin by the Peorian stage of deglaciation. The Geological Survey now (August, 1943) classifies the Iowan drift as representing an early substage of the Wisconsin stage, and the name Peorian is not used for either a distinct stage or a substage.

4. Correlation with European stages is merely suggested. Exact correlation is not justified by the information at hand.

5. Helen Jeanne Plummer collaborated by suggesting the names of several characteristic genera and species of Foraminifera.

6. The exact equivalence of the Port Hudson formation is unknown to the compiler.

7. The Vicksburg group has been restricted by the Geological Survey to include only the Marianna limestone and the Byram marl. The Glendon limestone is now classed as a member of the Byram.

8. The Fort Thompson formation is divisible into several members with the following supposed correlations:

Sangamon interglacial stage	FORT THOMPSON FORMATION	Coffee Mill Hammock marl
Illinoian glacial stage		Fresh-water limestone
Yarmouth interglacial stage		Marine-shell marl
Kansas glacial stage		Fresh-water marl
Aftonian interglacial stage		Marine-shell marl

9. The Key Largo limestone, the Miami oölite, and the Anastasia formation are contemporaneous facies. Surficial sands included by Cooke and Mossom (Florida, 1929) in the Anastasia formation are here excluded from it.

10. The Hawthorn formation as mapped by Cooke and Mossom (Florida, 1929) in Florida includes deposits as young as Choctawhatchee.

11. Vaughan (General, 1919, p. 203) reports from this horizon several species of corals closely related to species in the Rupelian of Italy.

12. The Black Mingo formation and the sandstone at Lillington, North Carolina, may be of Midway age.

13. The Pleistocene formations of New Jersey need restudy in the light of the glacial-control theory.

14. The Jacob sand and the Gardiners clay are put in the Yarmouth interglacial stage by Fuller (1914).

15. Fuller (New York, 1914) regards the Mannetto gravel as probably a glacial outwash, presumably Nebraskan. Wells (New York, 1935) thinks it is probably Pliocene. Its altitude above 270 feet presumably excludes it from the marine Pleistocene.

16. Fuller (New York, 1914) correlated the Manhasset formation with the Illinoian glacial stage. The correlation here used is essentially that of Wells (New York, 1935) and of MacClintock and Richards (New York, 1936).

17. Woodworth and Wigglesworth (New England, 1934) regard the Sankaty sand as of Nebraskan age.

18. Woodworth and Wigglesworth (New England, 1934) put the Weyquosque formation and the Dukes boulder bed in the Nebraskan.

19. The upper limit of the zone of large Lepidocyclinae, the zone of *Lepidocyclina gigas-L. undosa*, is moved upward to the top of the Oligocene to conform with the Atlantic and Gulf Coastal Plain part of the chart. For many years this zone—perhaps the most readily recognized Tertiary zone in tropical America—typified by the lower part of the Antigua formation of the island of Antigua, was assigned to the middle Oligocene, following Vaughan's (Caribbean, general, 1919a, p. 202–203) correlation of the accompanying coral fauna with that in the Rupelian (middle Oligocene) of Italy. The lower age limit of the zone is uncertain. It may represent a facies fauna appearing at almost any horizon in the Oligocene. At all events no well-defined lower Oligocene zone of orbitoidal Foraminifera is now known in the

West Indies and Central America. The only continuous Oligocene section in those regions for which a record of zoning on a basis of orbitoids is available is in Jamaica, where a zone of small Lepidocyclinae underlies the *Lepidocyclina gigas-L. undosa* zone (Matley, Jamaica, 1925).

20. As suggested when *Dictyoconus codon* was described (Woodring, Brown, and Burbank, Haiti, 1924, p. 608–609), that form is probably the microspheric form of a small *Dictyoconus* with which it is associated, *D. puilboreauensis*, a variety of which ranges upward into limestone of upper Eocene age. *D. puilboreauensis* is probably a synonym of *D. americanus*, based on material from the Saint Bartholomew limestone of the island of Saint Bartholomew, which was long considered upper Eocene but is now assigned to the middle Eocene. Nevertheless, large Dictyoconi of the type of *D. codon* have been found in middle Eocene formations in Jamaica and Haiti, but not in upper Eocene formations in any part of the West Indies. Whatever the relations of the different forms may be, large Dictyoconi of the type of *D. codon* so far as now known constitute a useful middle Eocene faunal zone in Jamaica and Haiti but are not recorded in Cuba. Small Dictyoconi are found in upper Eocene formations in Cuba, Jamaica, Haiti, and the Dominican Republic, and a large form having more numerous marginal partitions than *D. codon* is recorded from the upper Eocene of Jamaica (Vaughan, Jamaica, 1928, p. 282). The latest discussion of West Indian and Floridian Dictyoconi is by Cole (Caribbean, general, 1942, p. 12, 21–24), who unites *D. codon* and *D. puilboreauensis* with *D. americanus*.

21. The unnamed marl on Rio Yumuri, Matanzas, and the La Cruz marl in the Santiago de Cuba Basin, Oriente, are probably contemporaneous.

22. The name Yumuri limestone apparently should be suppressed in favor of the name Güines limestone. The Güines limestone of western and central Cuba, the Maquay formation of the Guantánamo Basin, Oriente, and the Manzanillo formation of the Cauto Valley, Oriente, are probably essentially contemporaneous. The Güines limestone and Manzanillo formation and correlated formations in Jamaica, Haiti, the Dominican Republic, Puerto Rico, St. Croix, and Florida contain the last species of the genus *Orthaulax*.

23. The Cobre and San Luis formations of southeastern and south-central Oriente, the Guaso limestone of the Guantánamo Basin, Oriente, and the Principe formation of central and western Cuba are probably essentially contemporaneous. *Discocyclina crassa* is not recorded from the Guaso limestone and Principe formation.

24. The Mancioneal formation is generally considered Pliocene. According to Trechmann's (Jamaica, 1930) account, its fauna is much like that of the Bowden formation, and the two formations may be of essentially the same age. Trechmann suggested that both might be considered Pliocene if they were in Europe.

25. The sandstone and conglomerate near Jacmel containing fossils considered of Pliocene age is underlain apparently without discontinuity by siltstone that may be upper Miocene. Fossils collected recently from the siltstone have not yet been identified.

26. The Miocene strata near Port-au-Prince are presumably Coryell and Rivero's (Haiti, 1940) Port-au-Prince formation, which was named without any description

of thickness and stratigraphic relations and without any locality, much less a type locality.

27. Foraminifera associated with *Discocyclina crassa* elsewhere, but not that species, are recorded from the upper Eocene of the Dominican Republic.

28. Olsson (Costa Rica, 1932, p. 42) assigned a longer time range to the Gatun formation of Costa Rica: late lower Miocene to upper Miocene.

29. The Emperador limestone was assigned to the upper Oligocene by Olsson (Costa Rica, 1932, p. 42); that is, he considered it older than the disputed upper Oligocene—early lower Miocene of the present chart.

30. The *Lepidocyclina gigas-L. undosa* zone has been recognized so far only in the Tonosi Valley, Los Santos.

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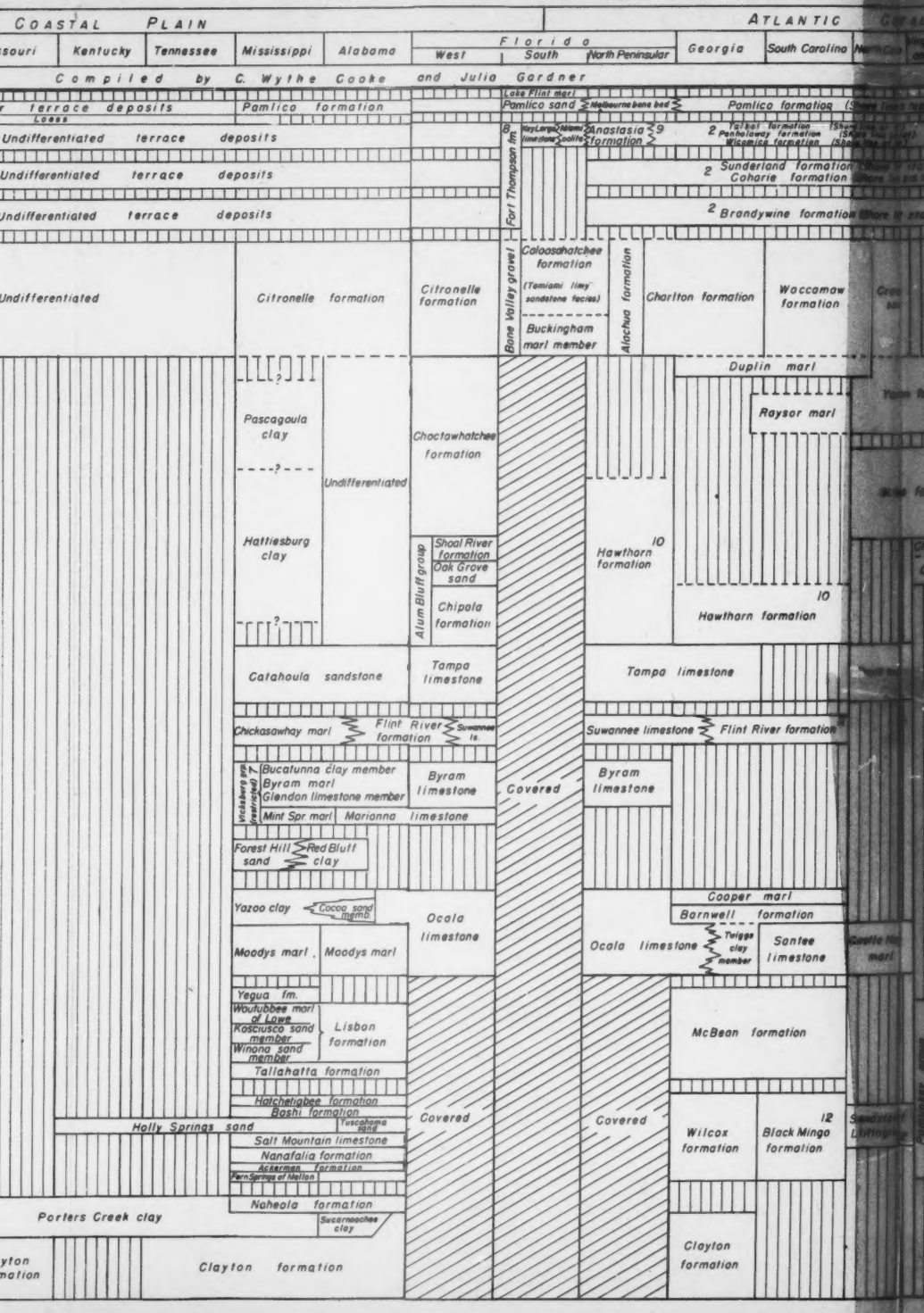
Chart No. 12. - CORRELATION

Generalized Standard Section				Faunal Zones	Gulf Coast					
	Series	American subdivisions	European[4] subdivisions	Charactistic fossils[5]	Northern Mexico	Texas	Louisiana	Arkansas	Missouri	K
										C o
Quaternary	Pleistocene	Wisconsin ("Peorian" Iowan)	Würm glacial	Smilodon floridanus	Undifferentiated		"Pensacola"	River terrac		
							Homeland terrace dep.	Loess		
		Sangamon	Würm–Riss interglacial		Undifferentiated	Beaumont clay	Port Hudson fm. & Prairie member ← (Fisk, 1938)	Undifferentia		
		Illinoian	Riss glacial							
		Yarmouth	Riss–Mindel interglacial		Undifferentiated	Lissie sand	Montgomery member (Fisk, 1938) Bentley member ← (Fisk, 1938)	Undifferentia		
		Kansan	Mindel glacial							
		Aftonian	Mindel–Günz interglacial		Undifferentiated		Williana member (Fisk, 1938)	Undifferentiat		
		Nebraskan	Günz glacial							
Tertiary	Pliocene	Upper	Astian		Undifferentiated	Willis sand (Pliocene?)	Citronelle formation	Undifferentia		
		Lower	Plaisancian	Anadara rustica Turritella peraltenuata Hipparion sp.		Goliad sand				
	Miocene	Upper	Sahelian: Pontian	Solenosteira vaughani Sconsia hodgei Turritella altilcostata Mulinia congesta Ecphora quadricostata Bulliopsis quadrata Turritella plebia Virgulina(Virgulinella) spp.	Lagarto clay (Miocene?)	Lagarto clay (Miocene?)	Pascagoula clay			
			Sahelian: Sarmatian							
		Middle	Tortonian	Valvulineria floridana	San Rafael beds	Oakville sandstone	Hattiesburg clay			
			Helvetian	Anadara strebla Chlamys nicholsi Turritella waltonensis						
		Lower	Burdigalian	Chlamys madisonia Anadara dodona Turritella subgrundifera						
				Orthaulax gabbi						
			Aquitanian	Turritella tampae		Catahoula sandstone				
	Oligocene	Upper	Chattian	Lepidocyclina favosa Chione bainbridgensis Cassidulus gouldii	Undifferentiated					
		Middle	Rupelian	Lepidocyclina supera Anadara lesueuri Lepidocyclina mantelli		Frio clay	Vicksburg limestone			
		Lower	Tongrian	Astarte triangulata Cardita carsonensis						
	Eocene	Jackson	Ludian	Discocyclina(Asterocyclina) spp. Camerina jacksonensis Aturia alabamensis Turritella alveata	Jackson formation	Jackson formation	Jackson formation	Jackson formation		
			Bartonian	Turritella arenicola Periarchus lyelli						
		Claiborne	Auversian	Ostrea sellaeformis Turritella nasuta	Yegua fm.	Yegua fm.	Yegua fm.	Claiborne group		
					Cook Mountain formation					
							Sparta sand			
			Lutetian	Lamarckina claibornensis Ostrea lisbonensis Turritella dutexata	Mt. Selman formation	Weches greensand mem. Queen City sand mem. Reklaw member	Cane River formation			
					Carrizo sand					
		Wilcox	Cuisian	Ostrea compressirostra Turritella humerosa Turritella mortoni	Wilcox formation	Indio fm.	Wilcox group	Wilcox formation		
			Ypresian	Cidaris splendens Ostrea thirsae						
	Paleocene	Midway	Thanetian	Vaginulinopsis echinata Volutocorbis limopsis	Midway group	Wills Point formation	Covered	Midway formation	Porters	
			Montian	Robulus pseudocostatus Hercoglossa ulrichi Ostrea pulaskensis Ostrea crenulimarginata		Kinkaid formation			Clayton formation	

TION OF THE CENOZOIC FORMATIONS OF THE ATLANTIC

By C. Wythe Cooke, Julia Gardner, and W. Th



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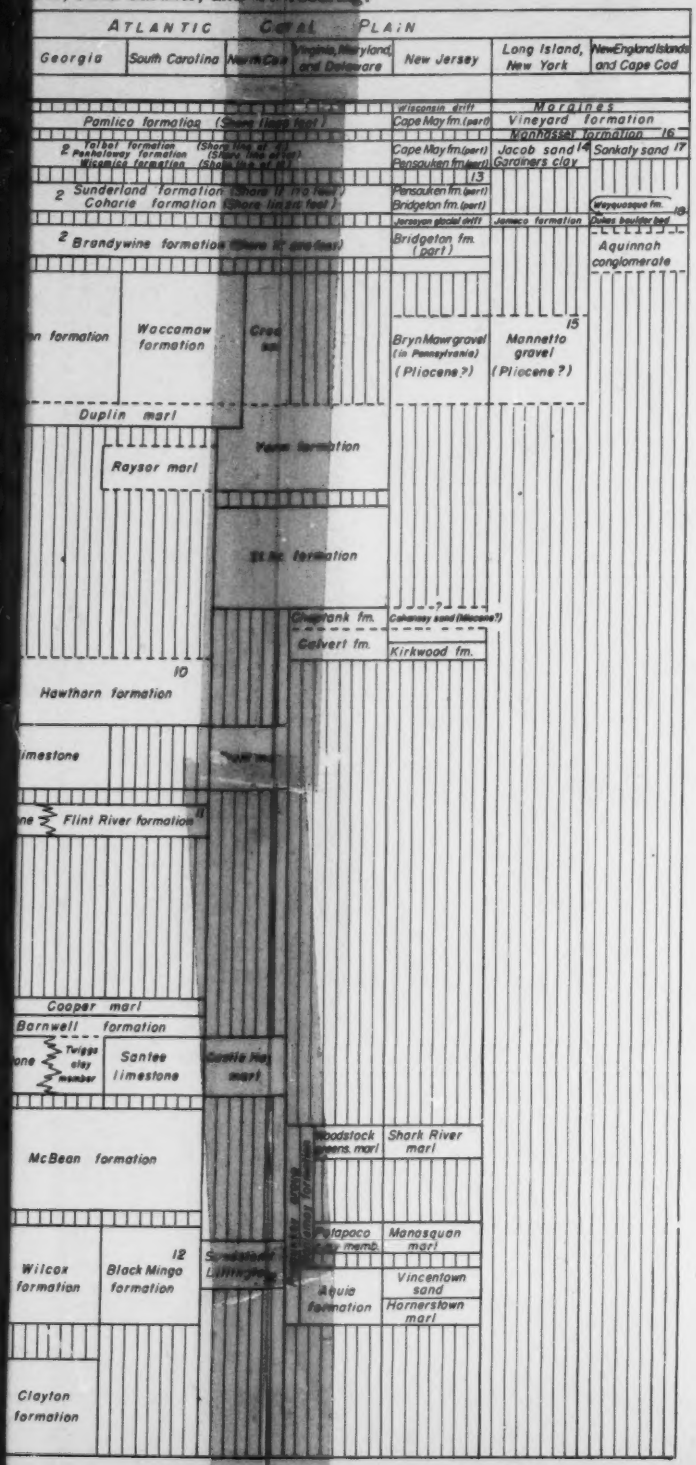

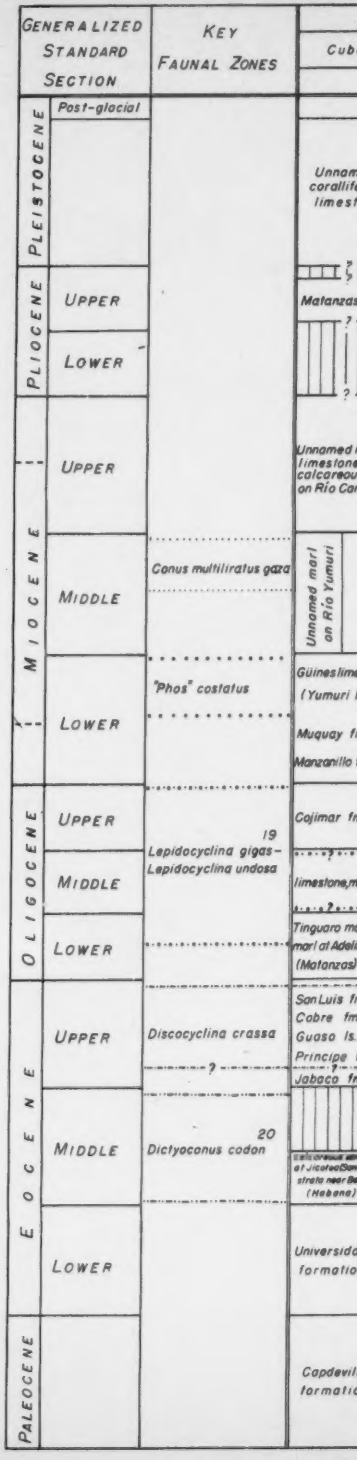

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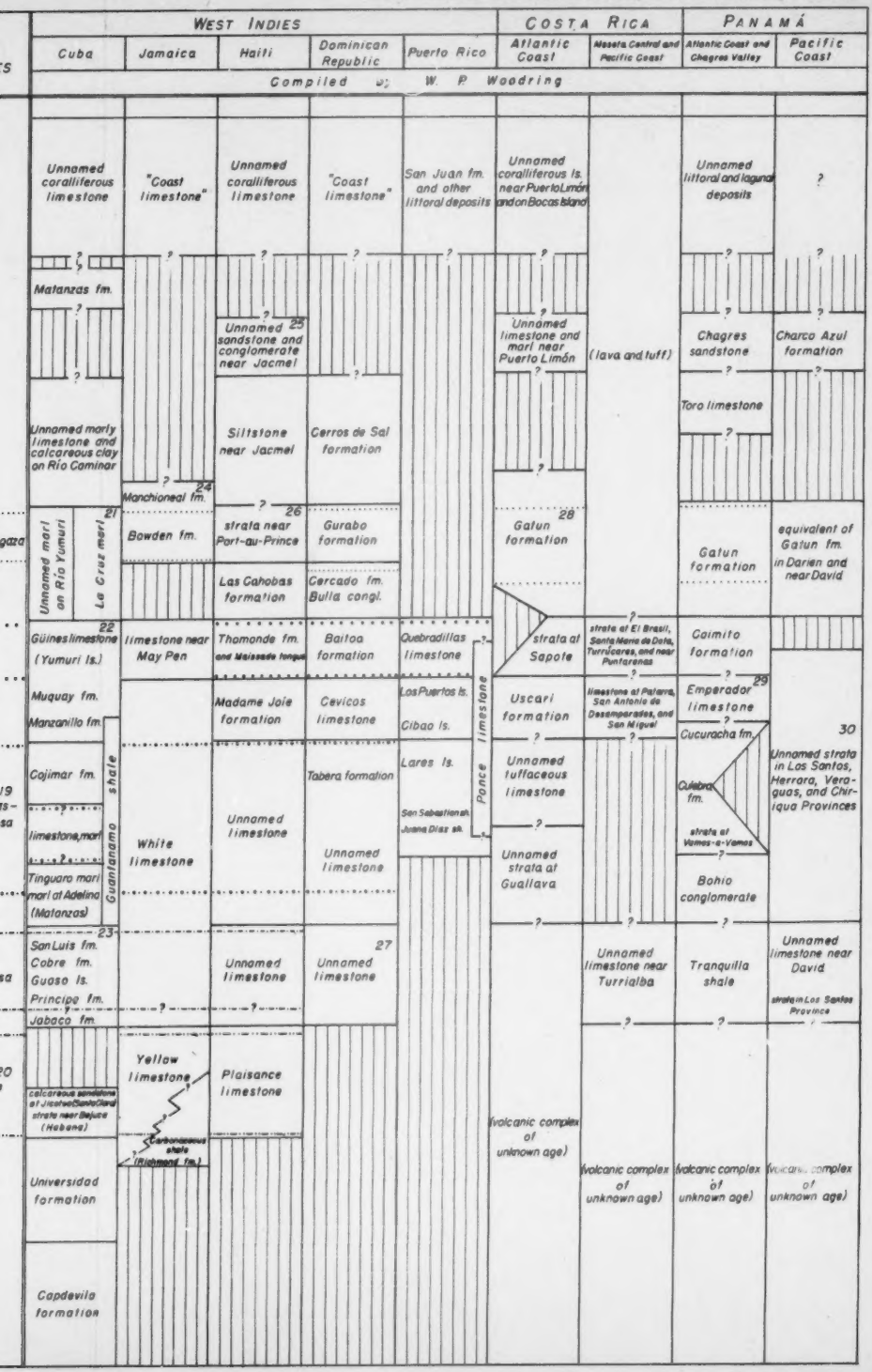
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